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Volume 23 JUNE, 1953 Number 2

Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
a Division of
The American Association of Petroleum Geologists

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Journal of Sedimentary Petrology

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NOTICE

Articles submitted for publication to the *Journal of Sedimentary Petrology* should have a prefatory abstract and sufficient postage should be sent by the author to pay the return of the manuscript as it has been found that many articles require more or less revision before publication.

Persons who are not members of the Society of Economic Paleontologists and Mineralogists or are not subscribers to the *Journal of Sedimentary Petrology* but who desire to publish an article in the *Journal,* should have some member of the Society of Economic Paleontologists and Mineralogists or some member of the American Association of Petroleum Geologists sponsor the article. Members of these Societies are given preference in publication.

Among papers to appear in forthcoming issues of the *Journal* are the following:

Some surface features of marine sediments made by animals. By K. O. Emery.

A study of porosity and grain relationships in experimental sands. By A. Gaither.

Tetrahedroid pebbles. By Daniel J. Jones.

Silicification in the Oneota dolomite. By N. Prokopovich.

The significance of evaporites. By L. L. Sloss.

Environment and mode of origin and facies relationships of carbonate concretions in shales. By L. G. Weeks.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, PP. 67–74
FIGS. 1, 2, PLATE 1, JUNE, 1953

PETROLOGY OF THE SPRABERRY SANDS OF WEST TEXAS

G. FREDERICK WARN AND RAYMOND SIDWELL
Texas Technological College, Lubbock, Texas

ABSTRACT

The Spraberry sands and siltstones consist largely of angular to subangular quartz, mostly of igneous origin. The sands are low in porosity and permeability because of close initial packing, pressurizing and tight cementation. The chief cementing material is opal with impurities of clay, carbonates and iron oxides. The most abundant inclusions in quartz are tourmaline, zircon and rutile. Authigenic and metasomatic minerals include dolomite, calcite, opal, chalcedony, quartz and manganese dioxide. Most of the carbonates have been replaced by chalcedony. Photomicrographs and micro-drawings illustrate these findings.

INTRODUCTION

The Permian Spraberry "sands" have a wide, proven distribution in the Midland basin portion of the West Texas-New Mexico Permian basin. The sands are mostly fine-grained and are interbedded with shales, laminated siltstones, silty sandstones, thin limestones, shaly limestones and dolomites. Most of the sand zones change to shales and/or limestones along the basin borders. The "trend" of the Spraberry formation appears to underlie a minimum of 1,000,000 acres distributed over parts of 12 or more West Texas counties. The present productive trend is approximately north-south and is 125 miles in length. The proven width is 50 odd miles and may eventually surpass 100 miles (fig. 1).

The Spraberry section averages 1,000 feet in thickness and is underlain by approximately 300 feet of fine-grained sands, dark shales and shaly limestones of the Dean sand. It is overlain by the shales and crystalline limestones of the Clearfork, or locally by the Tubb sand. The Spraberry is tentatively assigned to the lower Leonard, while the Dean sand is probably of upper Wolfcamp age (McLennan and Bradley). The Spraberry sediments were deposited slowly and intermittently in a subsiding basin. The basin occupied part of the Texas foreland and was surrounded by shelves and low platforms. The basin borders were mildly unstable. The clastic materials are of uncertain origin but may have been supplied by low-lying lands in the Ouachita-Marathon system south of the Midland basin (Warn, Carmack and Sidwell, 1952). There are two to three well-defined sand zones in the Spraberry section. Oil comes from both the upper and lower zones, with the former leading in production. The sands are closely packed, tightly cemented and are low in porosity, ranging from about 7 to 17 per cent. Most of the effective porosity is achieved through vertical fractures.

SPRABERRY LITHOLOGY

The lithologies of the Spraberry formation were described and classified by Gibson (1951). The major lithologies, in order of abundance, were given as shales, siltstones, limestones and sandstones. Gibson pointed out that most of the so-called sandstones actually are siltstones. On the basis of thin-section study he divided the sediments into four classes with the following characteristics: (1) Massive silty shale—65 to 78 per cent clay, 15 to 20 per cent carbonate grains, 5 to 10 per cent quartz grains and 2 to 5 per cent feldspar grains. (2) Black bedded shale—90 per cent clays and 10 per cent carbonate and quartz grains; partings parallel to the bedding. (3) Laminated siltstone—90 per cent quartz, 5 per cent carbonate and 5 per cent shale; laminations $\pm 1/16$ inch. (4) Siltstone, essentially a quartzite, 78 per cent angu-

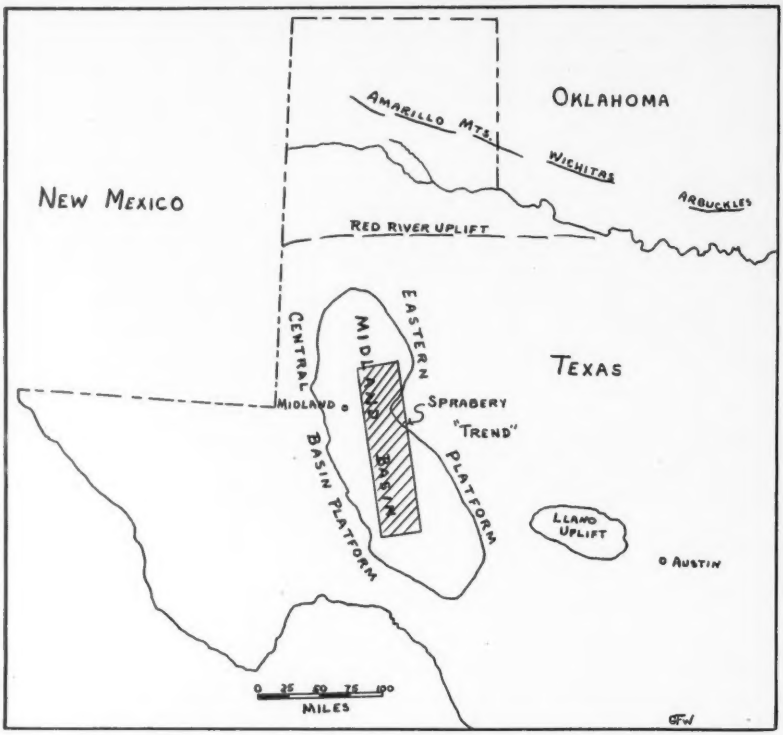

FIG. 1.—The Midland basin and approximate location of the Spraberry "trend."

lar quartz, 10 per cent carbonate grains and 2 per cent feldspar; permeability less than 1 md.

Ogden and Locke (1952) reported the following analyses from Spraberry cores in Midland, Glasscock, Reagan and Upton counties:

"Clean sand"—Permeability .002 to 2.5 md.; porosity 6 to 18 per cent; connate water 25 to 40 per cent; residual oil 15 to 25 per cent.

"Shaly sands"—Permeability .01 to 0.4 md.; porosity 6 to 12 per cent; connate water 25 to 90 per cent; residual oil 0 to 50 per cent.

Connate water is variable and tends to increase southward (Carlson, 1952).

SPRABERRY PETROLOGY

Petrology of the Spraberry sands was determined by the writers from a study of cores and of microscopic thin-sections from four of the cores. All of the cores were from fine-grained sandstones and silty sandstones and came from the following fields:

Tex-Harvey field	Midland County
George P. Livermore No. 1 Hughes	Reagan County
Sherrod field	Upton County
Pembrook field	Upton County
Benedum field	Upton County

Quartz.—The predominant mineral is quartz. Most of the quartz is clear to yellowish and rather well sorted. The

majority of the grains are angular to subangular. The grains are closely packed with numerous contacts showing the results of limited pressure and solution. About one-half of one per cent are rounded and subspherical, 3 to 5 per cent subrounded, 60 to 80 per cent subangular and 10 to 20 per cent angular. In the Sherrod field about 10 per cent of the medium and larger grains are subrounded and subspherical, and 25 to 30 per cent of the smaller grains are subrounded. Most of the grains are fractured and pitted.

From 50 to 80 per cent of the grains contain bubble and chain-like microlitic inclusions of gaseous and liquid origin. There are also inclusions of euhedral zircon, tourmaline and rutile. Approximately two per cent of the grains show strain shadows. The inclusions show that the quartz is predominantly of igneous origin. The grains obviously had a long history of physical and chemical wear prior to deposition as a clastic sediment. The angularity in fine-grained sands is ordinarily the result of impact during suspension-type transportation.

Chert is rare, occurring as subrounded grains with a diameter 20 times that of the largest quartz grain. The grain borders are strongly corroded. Chalcedony is rare as angular detrital grains. It occurs mainly as a secondary, replacing mineral in spines and in crinoid columnals and cirri (fig. 1, pl. I). It invariably shows well developed radial fibrous structure in thin-sections and is spherulitic in the columnals. The spherulitic cross is common under crossed nicols.

The cementing material of the sandstones consists mostly of opaline silica with clay, iron oxides and a small amount of calcium carbonate. Part of the opal has been derived from solution of quartz grains under static or mild diastrophic pressures.

Carbonates.—Carbonates comprise from one to three per cent of the rock volume and were most abundant in the cores from the Sherrod field. Most of the carbonates occur as small, irregular, scattered masses of impure calcium carbonate. Discoloring impurities include silt, clay, iron oxides and probably mixtures of ferrous and calcium carbonate. Some irregular masses show cleavage but no rhombs, and are doubtless detrital. Small scattered, isolated rhombs of discolored dolomite occur in several cores. The rhombs contain a central dark nucleus surrounded by an irregular carbonate growth, possibly calcite. The rhomb outline results from a second stage of carbonate growth apparently in the form of dolomite. Additional calcium carbonate is present as replacement material in organic structures, including algal fragments, columnals, cirri and spines.

Additional minerals.—Additional minerals include feldspar, garnet, tourmaline, zircon, rutile, natrolite, magnetite, hematite, limonite, pyrolusite, chlorite, muscovite and glauconite. Of these, only tourmaline, zircon and rutile are abundant, with pyrolusite locally abundant.

The feldspar occurs as angular, medium-sized grains of microcline with well developed cleavage. Some show little or no alteration while others are partially replaced by quartz. Garnet occurs as reddish brown fragments with altered edges. Tourmaline is light green, bluish green and brown, occurring as long slender euhedral crystals and crystal fragments, most of which are striated. Detrital fragments bearing the striations would tend to negate the possibility of prolonged transportation and abrasion for this mineral, unless suspension transportation was predominant. There are numerous inclusions of tourmaline in the quartz grains. Zircon is clear to light blue. It occurs as subangular fragments and tiny, short to elongate, prismatic subhedral to euhedral crystals, some of the latter being doubly terminated. Zircon inclusions in detrital quartz are numerous. Rutile is clear, yellowish and reddish brown and appears in long, delicate, thread-like and acicular crystals. Most of it occurs in detrital rutilated quartz.

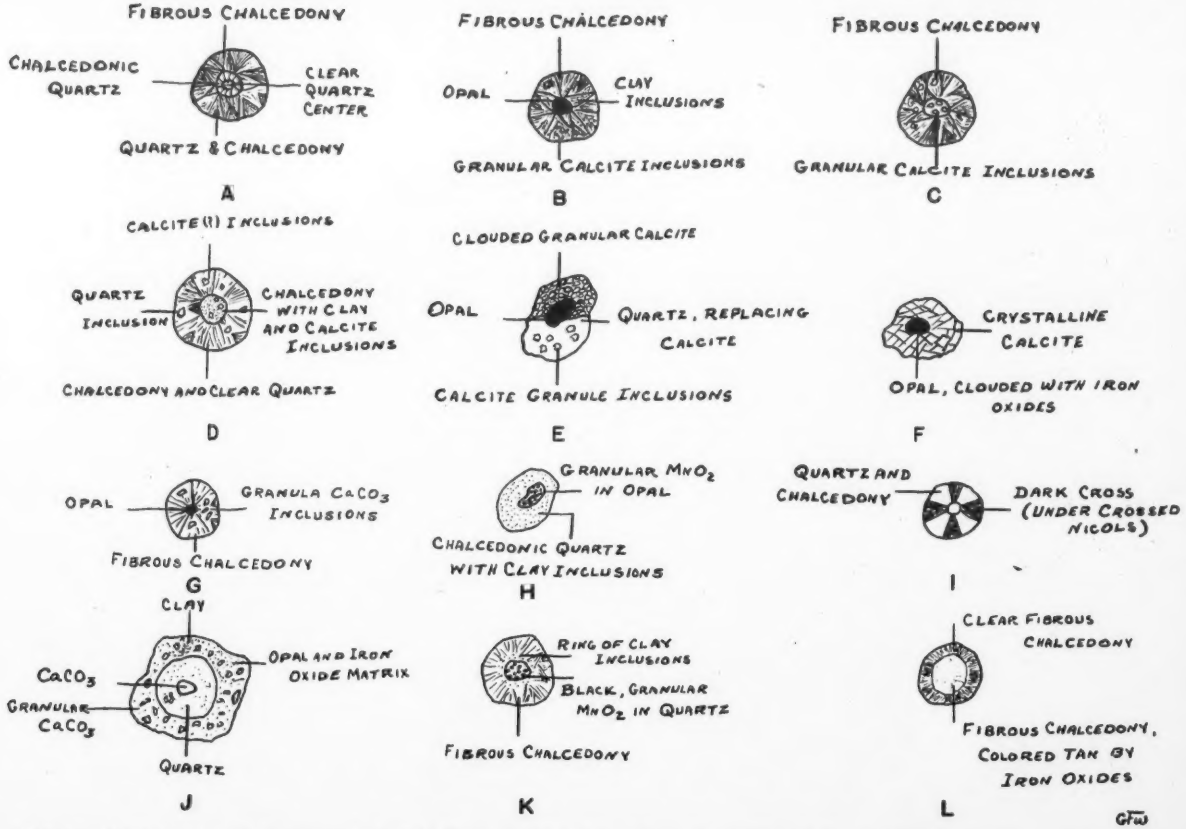

FIG. 2.—Micro-drawings of replacements in Crinoid columnals and cirri. They represent thin-section views under the polarizing microscope with the drawings enlarged from 6 to 8 diameters. The thin sections were prepared from core sections from the Pembrook and Sherrod fields. The chalcedony is light tan, becoming blue and fibrous under crossed nicols. The opal is opaque and has a deep bluish-black tinge under crossed nicols. Discolorations due to iron oxides and possibly iron carbonates are light to deep tan and translucent.

Magnetite occurs in the thin sections as unaltered black grains. The grains have approximately the same size range as the quartz. The grain forms are mostly subangular and are of detrital nature. Probably they were derived from igneous rocks.

Hematite is invariably reddish brown, appearing as scattered translucent flakes and irregular amorphous patches. Limonite is yellowish brown. It appears in oval or broken oval aggregates and as the alteration product of hematite. Diffuse limonite appears in some carbonates as well as in the opaline matrix.

Pyrolusite is black, appearing mainly in the form of small spherical aggregates and clusters of spherical aggregates. It occurs also in thin laminae composed of spherical aggregates. It is of sedimentary origin and here is classed as an authigene.

Chlorite is pale green, occurring in small, irregular flakes and is of metamorphic origin.

Muscovite is rare, appearing as colorless, elongate flakes with twinning.

Glauconite is pale green and pleochroic from yellowish green to green. It occurs in pellet form and is rare. The absence of glauconite penetration into the matrix suggests a detrital origin rather than an *in situ* origin for this mineral.

Organic materials.—Organic materials include spicules, spines, algal fragments, carbonized fragments, and columnals and cirri of crinoids.

The spicules are pale to yellowish green. They are long, slender and thread-like with pointed terminations. The spines are transparent, have needle-like form and thin double walls. They show some aragonitic structure indicating an original calcium carbonate composition. They are partially infilled by chalcedony.

The algal fragments are thread-like with wavy, silken aragonitic structure. They are preserved in calcium carbonate, presumably calcite, and show no effects of recrystallization or replacement. The carbonized material occurs in black, opaque amorphous patches. It appears granular and crumbly in contrast to the manganese dioxide aggregates.*

Columnals and cirri segments of crinoids occur only in the thin-sections from the Sherrod and Pembrook fields. Those from the Sherrod field show marked abrasional wear, indicating considerable transportation before deposition. Most of these fragments are replaced entirely by chalcedony (pl. I, fig. 1). The larger crinoid fragments from the Pembrook field are well preserved and have suffered little or no abrasion. The smaller cirri fragments were broken and rounded prior to deposition. They present a wide variety of replacement combinations, including crystalline calcite, quartz, chalcedony and opal. Chalcedony is the predominant replacing mineral (fig. 1).

CONCLUSIONS

The Spraberry sands accumulated in an elongate basin bordered by mildly unstable shelves and platforms. The sands appear to have been derived from relatively low-lying border-lands south of the Midland basin. The physical condition and small size of the quartz grains indicate a long history of reworking and transportation by water. The smallness, angularity and pitting of grains are typical of particles carried in water suspension. The paucity of unstable and metastable minerals associated with the quartz shows either that the parent rocks were thoroughly weathered in the southern borderlands or that the sands were derived from pre-existing sandstones in that area. The presence of interstitial clay, silt and iron oxide is not easily explained, unless one considers that they accumulated in mildly turbid waters where complete and final size sorting was impossible.

The original source rocks were mainly igneous. This is suggested by the numerous quartz grains containing inclu-

* Gibson (1951) reports that fossil leaves have been found in the Argo Oil Corp. No. 1 Cook well in Glasscock County.

under crossed nicols. Discolorations due to iron oxides and possibly iron carbonates are light to deep tan and translucent

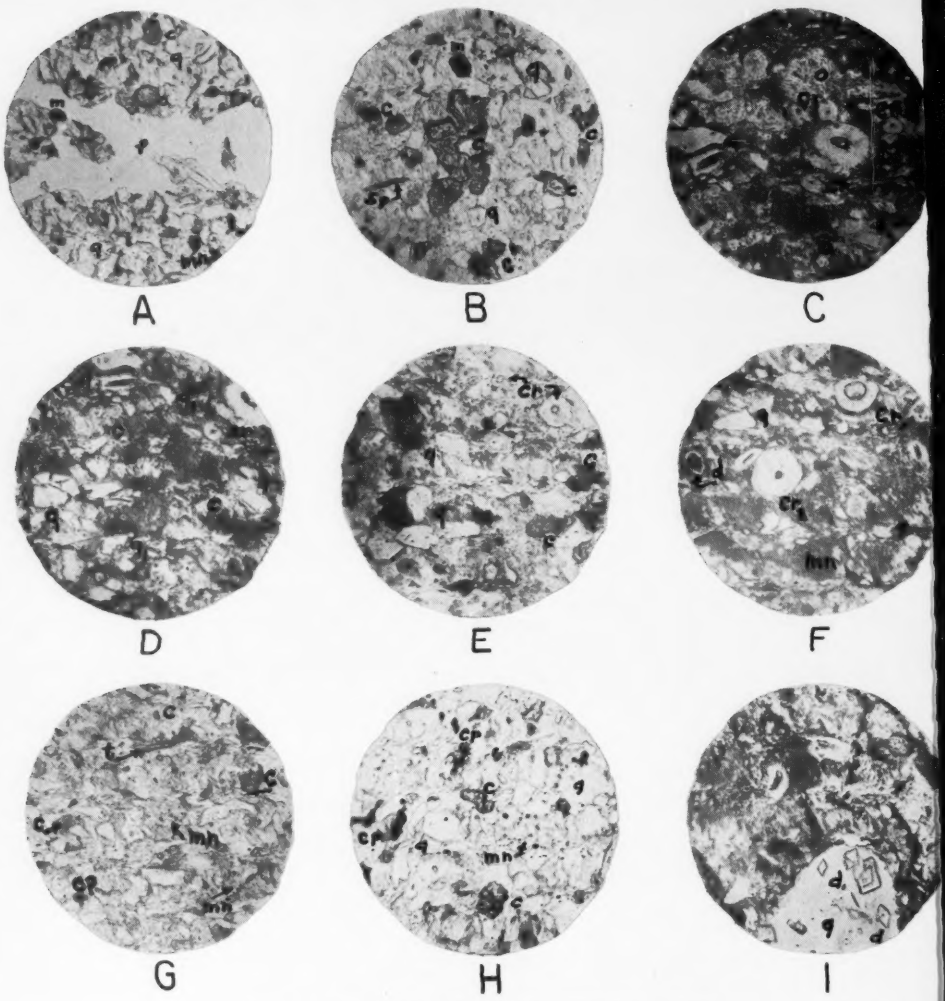

PLATE 1.—Photomicrographs of Spraberry thin sections.

sions of tourmaline, zircon and rutile. Detrital magnetite also suggests igneous source rocks. The rare occurrence of strain shadows in the quartz grains tends to eliminate metamorphic rocks as a primary source. The same can be said for the nearly complete lack of biotite, muscovite, chlorite and garnet.

Organic forms deposited with the sands include algae, echinoid spines,

LEGEND FOR PLATE 1

A—Sherrod field. Dense, very fine-grained silty sandstone with closely packed angular to subangular quartz grains (q). Approximately 50 per cent of the grains have microlitic inclusions. Most of the grain contacts are sutured or concavo-convex, showing the results of limited pressure and solution. The black subrounded grains are magnetite (m) and the dark, ragged, granular masses are secondary manganese dioxide (mn). The gray stippled clusters (c) are finely divided, cloudy calcium carbonate. Cement consists of small amounts of calcium carbonate, limonite and yellowish clay diffused in opaline silica. The bifurcate fracture (f) is completely filled by opal. Several triturated columnal fragments replaced by chalcedony occur in the slide but are not included in the photo.

B—Benedum field. Gray, dense, very fine-grained sandstone. Most of the quartz grains (q), are subangular; approximately 4 per cent are subrounded. Approximately 80 per cent contain microlitic inclusions. The grains are very closely packed and show limited pressurizing along the margins. There are no overgrowths. The granular clusters are calcium carbonates (c) with slight tan discoloration due to inclusions of clay and silt. The cement is opaline silica with diffused clay. The slender, thread-like object (sp) is an organic spine fragment partially infilled with chalcedonic quartz. The large black grain (m) is subrounded, unaltered magnetite.

C—Pembrook field. Laminated sandy siltstone with scattered grains of quartz and fragments of calcite, echinoderms and algae. Most of the quartz (q) is angular to subangular with partially corroded edges and a few thin overgrowths of silica. The circular to oval objects (cr) are crinoid columnals and cirri segments, replaced in most instances by chalcedony. The smaller segments show considerable abrasion, while the largest columnal has suffered marginal corrosion after replacement. The dark materials consist mainly of secondary manganese dioxide arranged in laminar fashion. The cementing material is chiefly opal (o) with small amounts of clay, calcite and iron oxide.

D—Pembrook field. Same thin section as C above. Angular quartz (q) with microlitic gas or liquid inclusions; quartz with corroded margins (q); granular calcite (c); columnals (cr); and same type of matrix as C. The angular grains on the far left probably result from partings along incipient fractures during compaction.

E—Pembrook field. Sandy siltstone with angular and subangular quartz (q) showing inclusions. Acicular inclusions are rutile. The columnals (cr) show effects of both corrasion and corrosion. The scattered aggregates of calcium carbonate (c) are sealed in by the opaline-clay matrix.

F—Pembrook field. Sandy laminated siltstone with scattered rounded to subangular quartz (q). The crinoid columnals (cr) show a minimum of abrasional wear and have been replaced by chalcedony. The large one in the upper right (cr) has a granular calcium carbonate center. The large one left of center (cr_2) is replaced with fibrous chalcedony and has an opal center. The dark areas contain granular and amorphous manganese dioxide (mn) and opal. The dark rhomb is a dolomite crystal (d) discolored tan by ferrous carbonate and is surrounded by a lighter, rhomb-like zone of secondary carbonate growth. The secondary growth is incomplete.

G—George Livermore No. 1 Hughes. Light gray, dense, very fine-grained sandstone. Most of the quartz grains are subangular to angular with about 5 per cent subrounded. Sixty per cent or more contain microlitic and bubble-like inclusions of probable igneous origin. The large majority of grains are fractured and pitted with partial impregnation by calcium carbonate and clay or iron oxides. The long slender fragment (t) is brown tourmaline. The small black dots are interstitial clusters of manganese dioxide (mn) comparable to pyrolusite. The crumbly black splotches (cp) are carbonized plant (?) fragments. Small, irregular aggregates of cloudy calcium carbonate (c) are scattered through the matrix. The cementing material is opaline silica with clay inclusions and a small amount of diffused calcium carbonate.

H—George Livermore No. 1 Hughes. Fine-grained quartzitic sandstone with angular to rounded quartz grains (q), scattered circular aggregates of manganese dioxide (mn), carbonized plant fragments (cp) and carbonate aggregates (c).

I—Pembrook field. Laminated sandy siltstone with dolomite rhombs (d) in quartz (q). The largest rhomb shows a small, clear, seed-like nucleus of calcium carbonate surrounded by an irregular, rhomb-like zone of carbonate growth. The final carbonate growth produced the outer, well defined dolomite rhomb. There are tiny calcite rhombs scattered in the matrix and included in some of the replaced columnals (not shown). The matrix is opal with finely divided calcium carbonate, iron oxides and clay. Manganese dioxide is arranged in very thin laminae.

The photomicrographs were taken with a polarizing microscope using a 10× eyepiece and a 10× (16 mm) objective. The photos were enlarged 1.8 times.

spicules, and crinoid fragments, all of marine origin. Most of these were benthonic forms and probably represent an appreciably wide depth range. The size discrepancy between the quartz grains and crinoid fragments shows that the latter were not transported far from the site of growth. This is especially applicable to the columnals showing little or no effect of trituration. Replacement of these forms by calcite, quartz, opal and chalcedony was both highly selective and variable.

Several lithologic features of the sands are responsible for low porosity and permeability: initial close packing of the grains; and interstitial silt, clay, and iron oxide. Partial solution of grain margins, further compaction, and deposition of resulting silica in the form of interstitial opal further decreased the porosity and permeability.

The calcium carbonate content is low, approximately one to three per cent by volume. Most of the original carbonate-bearing forms were replaced by quartz, opal and chalcedony. The remaining carbonates occur in scattered small patches which are effectively surrounded by siliceous cement. The closely knit rock fabric, plus the high silicate content tends to discourage a practical increase in effective porosity through ordinary acid treatment. However, the major portion of the cementing material is opal or impure opal. This material is more soluble in some acids than are the other forms of silica. It is possible, therefore, that selective leaching of the opal with special acids would increase effective porosity in some of the sand zones. Further petrologic studies should reveal means of obtaining improved reservoir performance and should aid in stratigraphic correlations. The origin of the sands may also be deciphered from their petrologic characteristics.

ACKNOWLEDGMENT

The writers are indebted to Roy P. Carmack of Stanolind Oil and Gas Company for use of some of the cores and thin sections.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, PP. 75–84
FIG. 1, JUNE 1953

ESTIMATION OF ERROR IN GRAIN SIZE ANALYSIS

J. C. GRIFFITHS
The Pennsylvania State College, State College, Pennsylvania

ABSTRACT

The estimation of error in grain size analysis and its comparison with the variation in size and size-sorting is reduced to an exact test of significance by means of analysis of variance technique.

The errors from a number of different rock types and environments are compared and the test is shown to offer a basis for judging the relative uniformity of a sediment in terms of size and size-sorting. The magnitude of the experimental error is considerably inflated by differences between operators and also, on the basis of the present evidence, it appears that there is a relationship between the magnitude of error and the mean size of the sediment.

ESTIMATION OF ERROR IN GRAIN SIZE ANALYSIS

The estimation of errors inherent in grain size analysis was first proposed by Wentworth (1926, 1927) and subsequently the sampling error was evaluated by Krumbein (1934); the last mentioned approach is of interest because of its general application to this and other experimental techniques. Krumbein indicated that there were many sources of variation in grain size analysis and strictly of course, there are an infinite number; the following items contributing to variation were specifically cited by Krumbein (op. cit., 1934):

1. Sampling error—is a function of
   (a) Homogeneity of the sediment
   (b) Locality collected
   (c) Kind of sample (e.g. one phase, composite, etc.)
2. Laboratory error—a function of
   (a) Subsampling error (e.g. sample splitting)
   (b) Sieving and all other operational errors

The term "error" is used here in its statistical sense meaning variation and does not imply mistake although it will certainly include variation due to "mistakes" as well as the other errors listed. Krumbein next reduced these errors to two large groups namely, variation associated with sampling an inhomogeneous sedimentary unit and variation associated with laboratory procedures of all kinds. As long as these errors or sources of variation are statistically independent an experiment may be specifically designed to segregate them and estimate their individual contributions.

The procedure advocated by Krumbein is to subdivide the samples by passing them through a sample splitter and to analyze an aliquot by sieving. This series contains variation from all sources or the variance of this series is an estimate of total variation. A second set of splits from the original samples is homogenized and this is considered to measure the sample to sample variation by apportioning it equally to each of the splits from the homogenized sample. The laboratory procedure is summarized in a flow sheet in figure 1.

The first series (A_{1-n}) is then a random sample of the original sediment. The statistics mean and variance (Snedecor, 1946) derived from the grain size analysis of each subsample of the set will be estimates of the variation in the population —the sedimentary unit. The second series (C_{1-k}) yields statistics which estimate the parameters of the homogenized population and any variation they show will arise from variation in laboratory procedure. The separate estimates of

TABLE 1.—*Mathematical model of analysis of variance for grain size analysis*

Source of variation	Degrees of freedom	Sums of squares	Mean square	F
Total (t)	$n-1$	$\sum_{1}^{n} X_A^2 - \frac{(\sum X_A)^2}{n}$	$\frac{ss_t}{n-1}$	$\frac{\text{m.s.}_t}{\text{m.s.}_e}$
Experimental error (e)	$k-1$	$\sum_{1}^{k} X_C^2 - \frac{(\sum X_C)^2}{k}$	$\frac{ss_e}{k-1}$	

variance are related through the equation:

$$s_t^2 = s_s^2 + s_e^2$$

where the variance s_t^2 is an estimate of the total variation and s_s^2 is an estimate of the sample variation, s_e^2 is an estimate of the experimental variation.

Now for equation 3 to hold it is necessary that each variance be reduced to an equal unit basis and this imposes the restriction that each subsample of the A and C set (see fig. 1) must contribute equal weight to the variance i.e. the samples must be equal to commence with; an important precaution is then to insure that each subsample of each set possesses an equal weight in grams as these are the units used in grain size analysis.[1]

The subsequent steps followed by Krumbein comprised converting the variance via the standard deviation to the probable error and the importance of the different sources of variation were qualitatively evaluated by comparing the contributions of s_s^2 and s_e^2 in terms of their probable error per cent (op. cit. 1934). It is clear that in order to detect valid differences in grain size between A subsamples it is necessary to show that s_e^2 is small compared with s_s^2. This is a rather inelegant termination to such a rigorous analysis and it appears to be more consistent to approach the solution to this last step with equal rigor to that of the earlier analysis. We may proceed as follows:

Returning to the variance equation we may divide through by s_e^2 yielding the equality:

$$\frac{s_t^2}{s_e^2} = \frac{s_s^2 + s_e^2}{s_e^2} \qquad (2)$$

Now either of these fractions is a variance ratio and the right hand side in particular is reminiscent of the variance ratio test used in analysis of variance by standard statistical procedure. Thus it is obvious that as the contribution from s_s^2 approaches zero the fraction $(s_s^2+s_e^2)/s_e^2$ approaches unity. To formulate this in statistical terms we may state that if the variation between samples is no greater than that arising from experimental analysis then the two estimates of variance are estimates of the same variance. This argument is the basis of the well-known F test (Snedecor, 1946, p. 218).

We may now arrange the computations and test of significance in an analysis of variance table (table 1).

The first column is self explanatory; columns 2, 3, and 4 are steps in calculation of the mean square or variance. The variance ratio (or F value) is obtained in column 5 by dividing the mean square for total by the mean square for error. The The significance level of the variance ratio may be obtained from standard tables of the F distribution (Fisher and

[1] I am indebted to Professor K. R. Bennett of the School of Agriculture, The Pennsylvania State College, for emphasizing this important precaution.

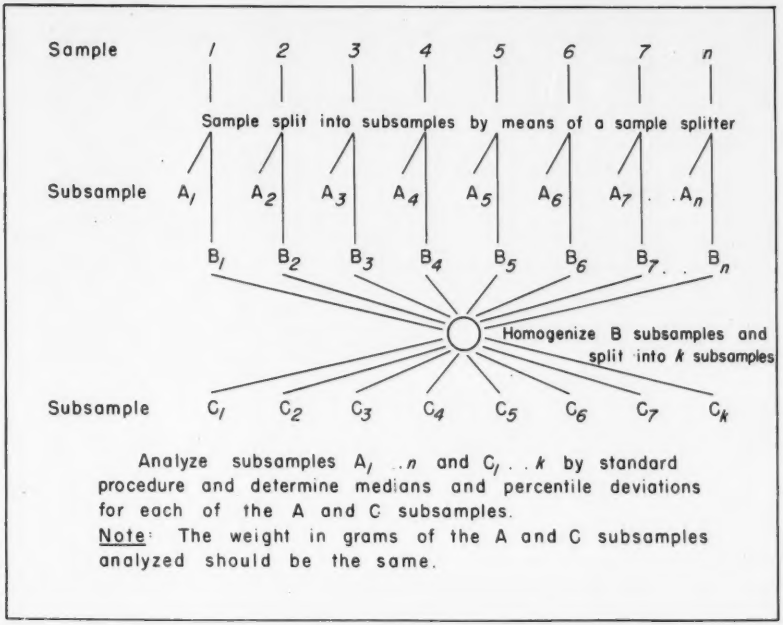

FIG. 1.—Flow Sheet and Instruction for Evaluation of Precision Error in Grain Size Analysis.

Yates, 1948, p. 34 ff., Snedecor, 1946, p. 222 ff.) using the degrees of freedom listed in column 2.

It may be of interest to enlarge upon the statistical argument leading to the significance test; we may first set up a null hypothesis (Fisher, 1942, p. 15) and if we use the median as the statistic characterizing the average grain size of each of our samples the null hypothesis becomes:

The samples are drawn from a homogeneous population and there is therefore no difference in median grain size between samples.

If this hypothesis were correct the variation we measure as variance is chance variation and the probability of obtaining a variance of any magnitude based solely on the number of degrees of freedom (number of samples minus one) is given by the corresponding F distribution. We may now compare our variation in terms of variance with the random variation (experimental error) and the F value obtained from this variance ratio is the basis for our decision. If the F value exceeds the tabulated F value at say the .05 (or 5%) point then such a difference would arise solely due to "errors" of random sampling less than 5 times in 100. We must decide therefore that either our hypothesis is true and this event is an unusual one (less frequent than one in twenty) or the variance ratio is larger than can be attributed solely to chance. In the latter case we reject our hypothesis and assume that there are real differences between the samples; or more correctly that the variation between the samples significantly exceeds that due to experimental error. If we decide to

reject the hypothesis we will on average be incorrect every time the rare event occurs, i.e. less than 5 times in 100.

This test has some interesting geological implications; thus suppose we find that the experimental error remains relatively constant then in several series of experiments the excess of the sample variance over the experimental error is a measure of uniformity of the sedimentary unit tested. The following examples may help to clarify the procedure and the geological meaning of the test.

EVALUATION OF PRECISION ERROR OF THE MEDIAN VALUE OF A POCONO GRAYWACKE

Six samples of Pocono low rank graywacke were collected from a bed of sandstone in an outcrop 4.2 miles north of Port Matilda on Pennsylvania highway 322. The experimental design followed the flow sheet of figure 1 modified to yield equal weights of six *A* subsamples and four *C* subsamples. The ten subsamples were analyzed for grain size distribution by sieving and pipette sedimentation and the phi medians and phi percantile deviations are listed in table 2.

The means and variances computed from the data are given in table 3 together with the standard deviations and coefficients of variation[1] for comparison with subsequent experimental results.

TABLE 2.—*Phi medians and phi percentile deviations of Pocono graywacke*

Sample no.	MD_ϕ	PD_ϕ	Operator[1]
	A Subsamples		
1	3.08	6.35	A
2	2.72	2.545	B
3	2.77	2.39	C
4	2.85	3.20	D
5	2.76	2.425	E
6	2.84	1.895	B
	C Subsamples		
1	2.72	5.35	A
2	2.90	3.59	C
3	2.90	2.25	D
4	2.82	2.345	E

[1] The column, operator, is inserted to indicate that the analyses were performed by five different operators distributed at random over the ten subsamples. Operator difference is therefore confounded with experimental procedure.

From the data in table 3 it is possible to compare the variances as suggested by Krumbein (op. cit., 1934), the coefficients of variation as suggested by Rittenhouse and Connaughton (1944) or to compute the variance ratio by means of the model of table 1. The last mentioned procedure yields the results given in table 4. The mean square is, of course, equivalent to the variance of table 3. We may now state our null hypothesis as:

The six samples of Pocono graywacke are all samples from an homogeneous population. Testing this assertion against the variation ascribed to experimental error we find that the variance ratio is less than the tabulated value at the 5 per cent level; in other words variations of such magnitude between medians of the six samples is likely to occur more than 5 per cent of the time by random sampling of a homogeneous population.[2] We accept the hypothesis and conclude that the sample differences are not significantly greater than the differences due to experimental error.

It should be clear from the above example that the experimental error contribution determines how small a difference we can detect in any particular investigation. In other words if we could reduce the experimental error we could establish valid differences between samples. Thus a comparison of the magnitude of error in analyzing different sediments should be informative.

Before commencing this comparison one other point should be stressed; one

[1] The coefficient of variation is defined as $Cv\,\% = \frac{s}{\bar{X}} \times 100$; where s = standard deviation = square root of variance.

[2] In fact from Fisher and Yate's tables (1948, pp. 37, 39) we can state that such differences are liable to arise on average less than 10 times but more than 5 in 100 such sets of samples.

TABLE 3.—*Statistics computed from the median and percentile deviations of the Pocono graywacke*

Statistic	A Samples		C Samples	
	MD_ϕ	PD_ϕ	MD_ϕ	PD_ϕ
$\overline{X}$	2.836	3.135	2.86	3.384
s^2	0.016668	2.65719	0.00213	2.091023
s	0.1291	1.6300	0.04618	1.44603
Cv. %	4.55		1.61	

TABLE 4.—*Analysis of variance of medians from Pocono graywacke*

Source of variation	D.F.	Sum of squares	Mean square	F	F_{05}	F_{01}
All sources	5	0.08334	0.016668	7.82	9.01	28.24
Experimental error	3	0.0064	0.0213			

of the assumptions at the basis of the *F* test is that the population should be normal and this we have insufficient evidence to decide in most cases but in practice considerable departures from normality are permissible (Cochran, 1947, Eisenhart, 1947, Fisher, 1942, Nissen and Ottestad, 1943). Nevertheless as the departure from normality becomes greater the probability levels become less exact and this is a feature to be remembered until we know more about the distribution of median grain size in sedimentary rocks. It is worth noting that had we been able to use means in the place of medians less importance would be attached to the above restrictions because means tend toward the normal distribution even when samples from which they are computed possess markedly non-normal distributions (Hoel, 1950, p. 68).

ESTIMATION OF PRECISION ERROR FOR SEDIMENTARY ROCKS

A number of results of such experiments compiled from the literature and from analyses carried out under the writer's direction are summarized in table 5. The median is the statistic tested in each case but whereas phi values were used in the writer's examples and phi values could be obtained from Krumbein's published grain size analyses of the Waverley Beach sand (op. cit.) the Rio Grande and Enoree River samples (Rittenhouse and Connaughton, op. cit.) were reported in millimeters and the absence of a mean value for the medians of the replicate C samples prevented a conversion to phi values in these cases. The *F* values for these two river sands were computed by using the coefficients of variation as estimates of variance and this is certainly not a recommended procedure.

We may now choose the 5 per cent level and state arbitrarily that any variance ratio which yields an *F* value not exceeding the 5 per cent *F* value of the corresponding *F* distribution will be considered to have failed to show significant differences. On this basis six of the 10 examples show significant differences, and in these cases valid differences in grain size as characterized by the median diameters can be established in the sedimentary unit sampled.

It is also informative to compare the coefficients of variation for experimental error;[1] five are less than 1 per cent and seven less than 2 per cent. The highest

[1] The advantage of the coefficient of variation depends on the fact that the Cv. is a pure number in which the effect of absolute size of the mean is removed.

TABLE 5.—*Estimation of error in*

Locality	Rock type or environment	Source of variation	D.F.	$\overline{X}_\phi$	s_ϕ^2	F
Waverley, Indiana	Beach sand	Total Error	23 7	1.922 1.934	0.00970 0.00012	80.8
Manzanilla, Trinidad, B.W.I.	Beach sand	Total Error	35 9	2.524 2.491	0.0687 0.0001	687.0
Lake Michigan Manitowoc, Wis.	Beach sand	Total Error	7 ?			1.33†
Miocene, S. Trinidad B.W.I.	4′ sand layer, low rank graywacke	Total Error	3 3	3.237 3.255	0.0123 0.0001	123.0
Miocene, S. Trinidad B.W.I.	33′ sand, low rank graywacke	Total Error	6 9	3.81 3.84	0.00233 0.00122	1.91
Rio Grande, New Mexico	River sand (fine)	Total Error	5 7	2.37 (2.37)	0.0081	32.1‡
Enoree River, South Carolina	River sand (coarse)	Total Error	23 7	−0.30	0.0081	9.25‡
Roslyn formation Washington State	Arkose	Total Error	15 7	1.2206 1.359	0.1647 0.0012	133.9
Gatesburg formation Central Pa.	"Quartzite	Total Error	15 6	2.21 2.19	0.0260 0.0631	0.41
Pocono sandstone Central Pa.	Low rank graywacke	Total Error	5 3	2.836 2.860	0.016668 0.00213	7.82

† With 7 degrees of freedom and an F value of 1.33 it is impossible even with infinite degrees of freedom for error to obtain a value of F_{05} less than 1.33; the F value was computed from the coefficients of variation which were in turn converted from Krumbein's probable error per cent.

‡ F values computed by treating the square of the respective coefficients of variation as estimates of variance.

value occurs in the Gatesburg experiment where 16 operators performed the 16 analyses. Differences between operators form a very large source in contributing to error. From other experimental investigations it has become obvious that it is essential to evaluate and effectively segregate operator error or the test will be relatively insensitive (Griffiths and Rosenfeld 1950, Rosenfeld and Griffiths 1950). It is also suggestive that the next two largest error contributions are associated with the coarsest sediments—Enoree river sand and Roslyn arkose respectively. If these results are maintained in future experiments it would appear that greater error will be associated with coarser sizes on average.

The succeeding step is to compare the total error after allowing for experimental error and this total error is clearly varying independently of experimental error so that it can be used to characterize the different kinds of sediments. The biggest total error in the tables is 33.25 per cent for the Roslyn arkose and this is indicative of the very great variability of the arkosic sediments in terms of median grain size. The Roslyn experimental error is not much larger than the others in the table and hence the major contribution to this variation is hetrogeneity in the sediment itself.

Two of the Trinidad experiments embrace two kinds of sand from the Miocene graywacke series of South Trinidad; the

determination of phi median diameters

F_{05}	F_{01}	F_{001}	Significance level P.	Cv. per cent	No. of operators	Source of data
3.44	6.15	12.73	<0.001	5.12 0.56	1	Krumbein (1934)
2.86	4.64	8.72	<0.001	10.3803 0.1204	2	Griffiths
			N.S.*	1.03 0.89	?	Krumbein (1934)
9.28	29.46	125.0	<0.01	3.43 0.003	2	Griffiths
3.37	5.80	11.13	N.S.*	1.26 0.91	2	Griffiths
3.97	7.46	16.21	<0.001	6.8 1.2	?	Rittenhouse & Connaughton (1944)
3.44	6.15	12.73	<0.01	7.3 2.4	?	Rittenhouse & Connaughton (1944)
3.52	6.35	13.71	<0.001	33.25 2.58	1	Bressler (1951)
3.96	7.60	17.99	N.S.*	7.29 11.46	16	Griffiths
9.01	28.24	134.6	>0.05<0.10	4.55 1.61	5	Griffiths

* N.S. = not significant.

Note: The values for the Manitowoc beach Sand, the Rio Grande and Enoree River sands are computed from the mean of the medians and the coefficient of variation, or probable error converted to coefficient of variation, and the statistics are quoted in the respective publications in terms of millimeter values.

first a four foot sand from which four samples were taken and four replicates run. The error is fortuitously low and the differences are significant for this reason but the total variation exceeds the variation in the second Trinidad example. In other words, irrespective of experimental error, one would expect a 33 foot sand body to show much greater variation than a four foot sand of the same series. The 33 foot sand is, of course, unusual and its uniformity has been established on other grounds (Griffiths, 1951, p. 233).

The Gatesburg quartzite despite the large experimental error possesses a relatively low total coefficient of variation. This suggests that the Gatesburg quartzite is much more uniform than the Roslyn arkose in terms of median grain size—a feature, of course, which confirms our expectations. It is, however, somewhat disappointing to note that the beach sands, river sands, quartzites and low rank graywackes show variation of the same order of magnitude. Presumably this indicates that we need a much larger number of samples in order to elucidate such differences. It may well be, however, that the measures of central tendency such as median and quartile deviation are not sufficiently sensitive for our purpose because there is little doubt that fluvial sorting is as efficient as beach sorting in terms of the sand sizes and the central part of the size distributions are therefore likely to be similar; the gray-

TABLE 6.—*Precision error of the phi percentile deviation*

Locality	Source of variation	D.F.	X_ϕ	s_ϕ	s_ϕ^2	F	P	Cv per cent
Waverley beach sand	All sources	23	0.489	0.074094	0.005490	7.02	<.01>.001	15.15
	Experimental error	7	0.209	0.027964	0.000782			13.37
Manzanilla beach sand	All sources	35	0.572	0.193675	0.037505	202.73	<.001	33.86
	Experimental error	9	0.613	0.013601	0.000185			2.22
Gatesburg quartzite	All sources	15	4.16	1.520661	2.31240	3.27	N. Sig. $P<.10>.05$	36.55
	Experimental error	6	4.09	0.841348	0.707866			20.57
Pocono graywacke	All sources	5	3.134	1.6300	2.657054	1.27	N. Sig. $P>.20$	52.01
	Experimental error	3	3.384	1.44603	2.091023			42.73
Miocene 4′ sand	All sources	3	1.565	0.175973	0.03096	10.567	<.05>.01	11.24
	Experimental error	3	1.660	0.017127	0.00293			0.10
Miocene 33′ sand	All sources	6	1.543	0.279028	0.077856	1.36	N. Sig. $P>.20$	18.08
	Experimental error	9	1.721	0.238954	0.057099			13.88
Roslyn arkose	All sources	14	1.906	0.839270	0.704368	91.24	<<.001	44.03
	Experimental error	7	1.614	0.087864	0.007720			5.44

wackes may only be differentiated from beach and river sands by presence of clay sizes in the former and its absence in the latter cases.

TESTING OTHER CHARACTERISTICS OF GRAIN SIZE DISTRIBUTION

Similar tests were performed upon several of the sediments in terms of the phi percentile deviation (Griffiths, 1951), a measure of sorting, and the results are summarized in table 6. Four of the seven examples yielded significant differences and in three of these, Manzanilla beach sand, the four foot Miocene sand, and in the arkose of the Roslyn formation, experimental error as measured by the coefficient of variation is very much less than the variation between samples. In the case of the Waverley beach sands the error forms a large part of the total variance.

In each one of the non-significant examples experimental error accounts for a very large proportion of the total variation. In two of the examples, the Gatesburg and Pocono experiments, this is very likely to arise from the differences between operators, whereas in the Trinidad Miocene 33-foot sand it is essentially a reflection of uniformity of the sediment.

In using the phi percentile deviation one source of variation is the reading error which becomes large when the percentile intersection occurs in a relatively flat portion of the cumulative curve. In most of the examples given above this factor is of considerable importance because the 90 percentile falls in the pipette portion (silt fraction) of the curve. In one case, the Gatesburg, it is clear that the site sampled is not typical of Gatesburg orthoquartzite because the samples were taken from the weathered portion at the top of a sand pit. The clay and silt is presumably a weathering product not related to original Gatesburg deposition. With this in mind we may calculate the size distribution of the Gatesburg samples in terms of 100 per cent greater than 62.5 microns i.e. 100 per cent sand fraction. The results of this calculation are given in terms of median and percentile deviation in the first two rows of table 7.

The differences between samples in terms of median diameters are still not significantly greater than the differences due to experimental procedure but the magnitude of total variation has been very markedly reduced. The error as measured by the coefficient of variation is now of the same order as that of the Roslyn arkose. In the case of the phi percentile deviation the differences be-

TABLE 7.—*Precision error in Gatesburg quartzite*

Locality	Source of variation	D.F.	X_ϕ	s_ϕ	s_ϕ^2	F	P	Cv. per cent
100% sand MD_ϕ	All sources	15	1.949	0.137785	0.018985	2.50	N. Sig. $P<.20>.10$	7.07
	Experimental error	6	1.936	0.087149	0.007595			4.50
100% sand PD_ϕ	All sources	15	1.04	0.134830	0.018176	5.45	$<.05>.01$	12.96
	Experimental error	6	1.01	0.057732	0.003333			5.72
Amount at 8_ϕ	All sources	14	86.66	4.42580	19.588100	1.97	N. Sig. $P>.20$	5.11
	Experimental error	6	86.93	3.15580	9.959185			3.63
Amount between 5–9_ϕ	All sources	14	7.30	4.370	19.10	4.937	$<.05>.01$	59.86
	Experimental error	6	7.57	1.967	3.8690			25.98

tween samples are significantly greater than the differences due to experimental error at the 5 per cent level and again the magnitude of error variation has been markedly lowered.

To further test this aspect of the technique the amount of Gatesburg sediment in terms of per cent by weight were read from the curves at the 8_ϕ (3.9 micron) diameter ordinate. The variation between samples is not significantly different from the variation due to experimental error but the magnitude of variation (Cv. %) is quite small. Apparently both pipette and sieving lead to errors of similar magnitude in terms of median (100% sand) and per cent by weight at the 8_ϕ diameter.

The per cent by weight of sediment between 5 and 9_ϕ (31.2 to 1.78 microns respectively) once again yields significant differences at the 5 per cent level. The magnitude of variation is now however, extremely large both in terms of total or experimental coefficients of variation.

It appears from this preliminary examination that whereas sieving shows a similar degree of experimental control to that of pipette sedimentation in terms of single points on the curves, spreads are much more variable and particularly in the region covered by pipette sedimentation. It is difficult under average laboratory conditions to standardize pipette sedimentation and, of course, very much more difficult if a large number of operators take part in the experiment. There is little doubt that the largest source of error in the phi percentile deviation is due to errors in drawing and reading the cumulative curve in the region of the 90 percentile which in most natural rocks is found in a flat part of the curve. This suggests that the 90 percentile is rather sensitive and should be used with caution in any measure of spread and that any statistic dependent on an ordinate beyond the 90 percentile will suffer somewhat greater variability than the 90 percentile.

This leads to something of a dilemma; in selecting a statistic as a measure of sorting the open ended size distributions compel us to adopt quartile or percentile measures. Using the phi quartile deviation (or Trask's sorting coefficient) it is not possible to differentiate between delta sediments and beach sands in terms of sorting and the quartile measures are thus too insensitive. On the other hand in those sandstones with clay factions the ninety percentile may be reflecting error rather than "true" variation in many cases hence the percentile measure is too sensitive. It is, of course, possible to choose some other pair of percentiles between the quartiles and the ninety and ten percentiles but this should be a last resort because the main usefulness of a statistic is achieved when it is derived from a well-known distribution function and generally the more arbitrary the choice of statistic the less efficient the statistic becomes. In this respect it may be well to consider very carefully whether our techniques—which ultimately give

rise to these difficulties—are really yielding the most meaningful measure of size?

ACKNOWLEDGMENTS

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FIGS. 1–5, JUNE, 1953

BEACHROCK IN SOUTH FLORIDA[1]

R. N. GINSBURG
Marine Laboratory, University of Miami, Coral Gables, Florida

ABSTRACT

The rapid intertidal lithification of beach deposits in the coral seas has received the attention of numerous investigators. Study of beachrock from the Dry Tortugas, Florida, shows that the aragonite cement is precipitated from the sea water remaining in the beach sands at low tide. High temperatures, rate of beach drainage, and the permanence of the beach, control the localization of beachrock. The recognition of beachrock in the fossil record is briefly discussed.

INTRODUCTION

Beachrock or beach sandstone consists of calcareous skeletal debris cemented by calcium carbonate to form a friable to well-cemented rock; where mineral grains and rock fragments are common in the beach deposits they also occur in the beachrock. The particles are mostly sand-sized with beds composed of reef shingle or conglomerate, and intraformational breccia (Kuenen, 1933, p. 87) is less abundant. Recent beachrock occurs discontinuously in the intertidal zone, on islands with sand beaches, as a series of thin beds which dip seaward at less than 15° and whose strike is parallel to, or at slight angles with, the trend of the present beach. Exposed beachrock is usually broken into rough rectangular blocks by joints parallel and perpendicular to its strike, and in some areas these blocks lie below low tide with a steeper than ordinary dip. The maximum thickness of recent beachrock is probably not more than the order of the eight feet reported by Vaughan (1914, p. 63). Kuenen (1933, p. 87) calls attention to the " . . . very abrupt, flat junction between the beachrock and the substratum."

The formation of marine beachrock is limited to the latitudes of coral growth. Kuenen (1933, pp. 86–87) refers to descriptions of beachrock from the East Indies and adjacent regions. Other observations and descriptions may be found in Gardiner (1902, 1906, 1930), Branner (1904), Sewell (1935, 1936) Steers (1940), and Fairbridge (1948, 1950). The most detailed account of beachrock is that of Branner (1904) who studied it along the northeast coast of Brazil where it occurs intermittently for a distance of some 1,000 miles. Daly (1924, p. 139), Kuenen (1950, p. 434), Fairbridge (1948, p. 74), and others have called attention to the rapidity of beachrock formation.

The purpose of this paper is to describe the only occurrence of marine beachrock easily accessible from continental United States, and to discuss the formation and distribution of this rock.

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BEACHROCK IN FLORIDA

Typical beachrock has been found in Southeast Florida only in the Dry Tortugas. Vaughan (1914, p. 63) and Field (1919, p. 198) mention its occurrence around Loggerhead Key. It is found on both sides of the narrow island as shown in figure 1. The strike is generally parallel to that of the present beach though differences as large as 15° have been ob-

[1] Contribution no. 92, Marine Laboratory, University of Miami.

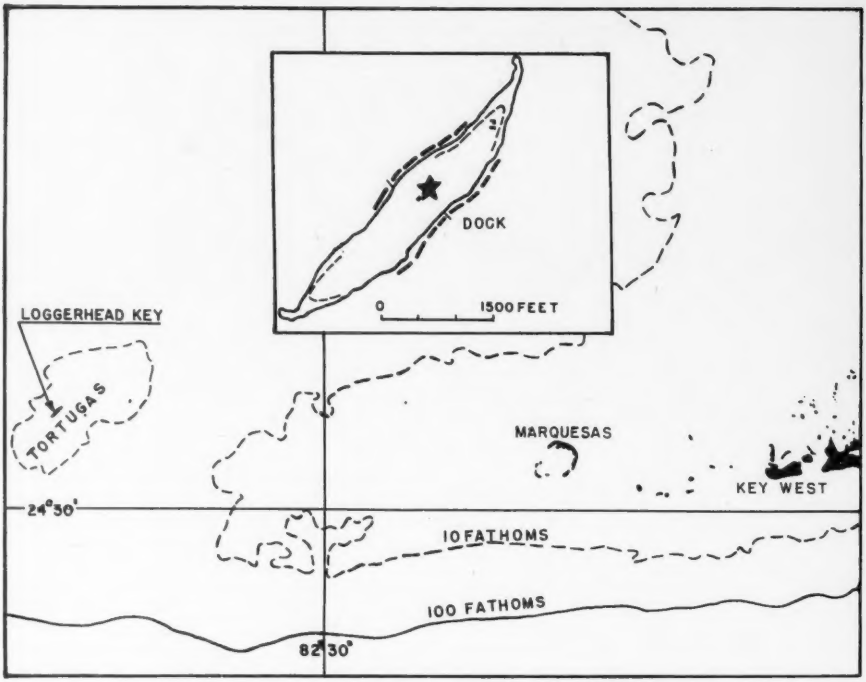

FIG. 1.—Chart showing location of the Dry Tortugas and Loggerhead Key. Insert is an outline map of Loggerhead Key after Vaughan (1914) on which the distribution of beachrock is shown by the heavy dashed lines.

served. The dips do not exceed 10°. As shown in figure 2, the truncated ends of the thin beds form a series of small cuesta-like ridges. Vaughan (1914) reported eight feet or more of beachrock at this locality.

Some exposures have smooth, nearly planed surfaces, often with small circular pits which are similar to the "stonelace" described by Emery (1946, p. 227). More typically the rock has a pitted, fissured, and honey-combed appearance, figure 3, which is characteristic of shore-exposed calcareous rocks in the coral seas. Here the rock is very brittle and its surface is discolored by algae.

Petrography

The beachrock is a thin-bedded, well sorted calcarenite with some beds of coral conglomerate. In hand specimen the grain size distribution within each bed appears uniform, but differences between adjacent beds produce distinct textures and fabrics. Variations in grain size on a finer scale are found within the beds of sand size debris giving laminae of the order of a centimeter in thickness, as shown in figure 4, and as described by Emery (1950).

The grains are fragments of algae, coral, molluscs, foraminifera, bryozoans, etc. Some of this skeletal debris in the

FIG. 2.—Beachrock on the northwest side of Loggerhead Key showing the seaward-dipping thin beds.

beachrock still retains its original tint, and X-ray spectrographic analyses made through the courtesy of H. A. Lowenstam show that the content of aragonite is 90 and 92.5 per cent, respectively, for two samples as compared with 87.5 and 90 per cent for two samples of beach sand from the same locality. The unaltered character of the skeletal debris is confirmed by thin-section examination; grains of *Halimeda*, coral, molluscs, and incrusting algae showed no recrystallization or selective solution.

Considerable variation in the degree of cementation can be seen even in hand specimens. Some of the beds of medium to coarse sand-sized debris are so poorly cemented that they can be reduced to their component grains by rubbing with the fingers; other beds, generally but not always, those of finer grain size, are well cemented though still porous (see fig. 4.). Some idea of the cementation can be gained by comparing the porosity of the beachrock with that of beach sands. The porosity of four samples of beachrock, whose cementation can be described as ranging from fair to moderate, varied from 30 to 38 per cent, with an average value of 34 per cent. Daly (1924, p. 136) gave values of 14.7, 14.8, and 16.5 per cent for the " . . . more thoroughly indurated beachrock . . . " from the east side of Loggerhead Key. He indicated that the porosity of average beachrock was higher than these values, and that recently formed beachrock had a porosity of the order of 35 per cent. Beach sand from the present beach on Loggerhead Key had a porosity of 42 per cent when compacted to minimum volume by tapping *In situ* porosity of a beach sand, similar to that of Loggerhead Key, collected on the lower Florida Keys, was 55 per cent for two samples.

The nature of the cementation is

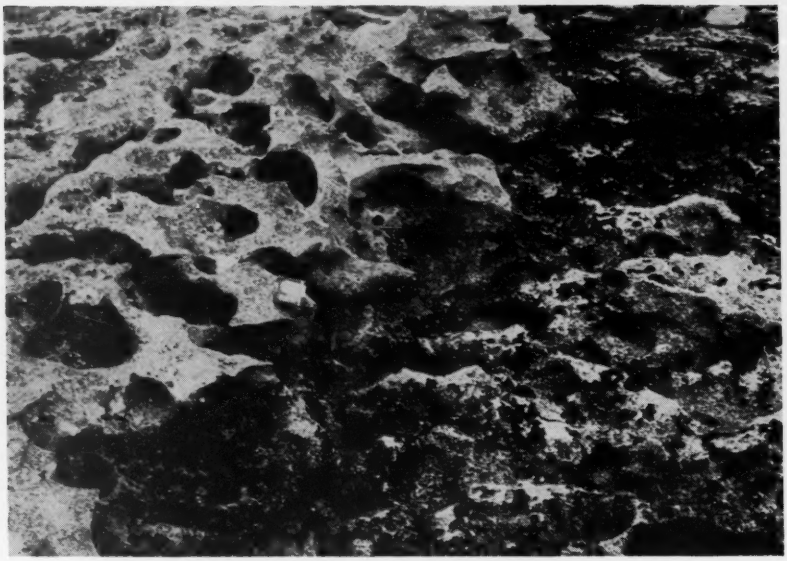

FIG. 3.—Intertidal erosion of beachrock northwest side of Loggerhead Key. Approx. 1/20 ×.

clearly shown in thin-section; around almost every grain is a fringe of acicular crystals of calcium carbonate, whose long axes are perpendicular to the grain border (see fig. 5). These crystals are from 0.005 mm to 0.04 mm in length and are stained lilac by cobalt nitrate, identifying them as aragonite. The borders of the detrital grains are usually brownish though the grain is clear, and the fringes of aragonite are also a dusty brown, suggesting iron oxides.

ORIGIN OF BEACHROCK

The two main hypotheses of beachrock formation—the action of fresh ground water, and that of sea water—have been discussed by Branner (1904, p. 196), Daly (1924, p. 135), Kuenen (1933, p. 87) and others. Solution of calcium carbonate from the island sands by ground water, followed by precipitation as this water seeps out through the beach at low tide was suggested by Field (1919, p. 198). Daly (1924, p. 138) has objected to this view because it fails to explain the sporadic occurrence of beachrock, and the cementation of deposits low in detrital calcium carbonate. Daly (1924, p. 139) believes that rapidly deposited sands, as from a storm, are required for beachrock formation. Bacterial decomposition of an assumed higher than usual content of organic matter in these deposits is thought to produce the precipitation of an initial aragonite cement which fixes the sand. Inorganic precipitation from sea water then accomplishes the major cementation

Gardiner, Kuenen, Daly, and others interpret the localization of beachrock in the intertidal zone as evidence that the cementation is due to the precipitation of calcium carbonate from sea water, primarily as a result of heating and evaporation. The nature of the cementation supports this interpretation.

Recent investigations of the apparent

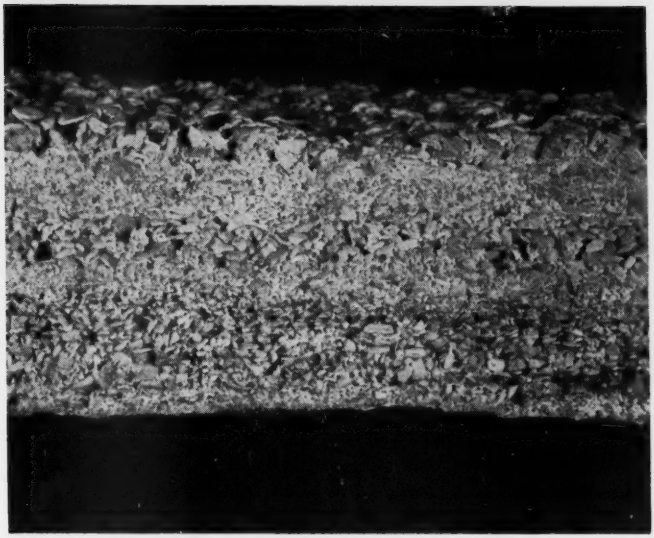

FIG. 4.—Vertical section of beachrock ×2. The two laminae of finer grain size show better cementation than the adjacent coarser laminae.

solubility product of calcium carbonate in sea water, by Wattenberg and Timmerman (1936), Smith (1941), and Hindman (1943), indicate that normal surface sea water is supersaturated with respect to this component, though there is some difference in the results regarding the extent of supersaturation. These workers suggested that the absence of nuclei is an important factor in the maintenance of this apparently anomalous supersaturation.

Beach sands provide abundant nuclei for the precipitation of calcium carbonate from water contained within them. Moreover, the following factors, which are important in the intertidal zone of beaches, will aid the precipitation of calcium carbonate: (1) Increased temperatures within beachsands promote precipitation because the apparent solubility product decreases with increasing temperature, and because of an increase in the partial pressure of carbon dioxide. Longley (1917, p. 185) found the temperature of the beach surface at low tide on Loggerhead Key to be 39° at 1445 hrs., August 10, 1917. (2) The water remaining in the beach at low tide after drainage is in the form of intergranular films. This condition allows more complete exchange of carbon dioxide between atmosphere and solution, making possible more rapid equilibration; i.e., precipitation from supersaturated solution. (3) Interstitial precipitation may be aided by evaporation of the residual films of water at low tide.

DISCUSSION

The absence of any grain solution within the beachrock indicates that the aragonite cement has not been derived by internal solution and reprecipitation. Additional evidence that the cement is precipitated from sea water is beachrock which contains little or no detrital calcium carbonate (Branner, 1904, Daly, 1924).

On the view that normal surface sea

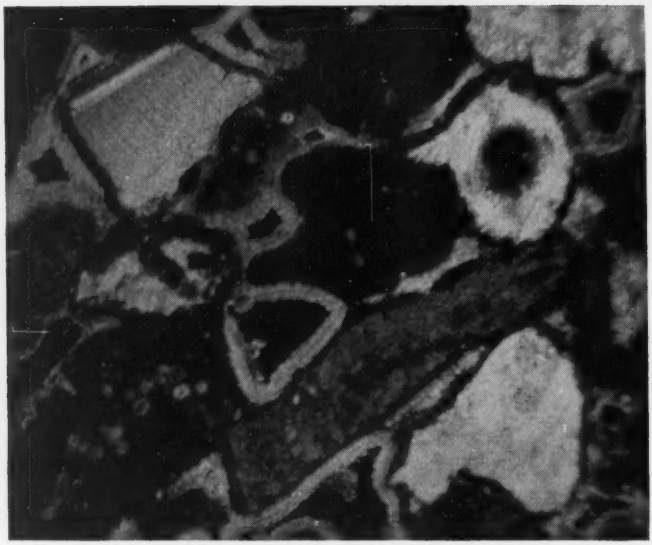

FIG. 5.—Thin section of beachrock, polarized light, ×47. The white pore linings and grain borders are fringes of acicular aragonite.

water is just saturated with respect to calcium carbonate, concentration by evaporation is necessary for precipitation. Emery and Foster (1948) found that evaporation from the intertidal beach in Southern California is confined to the upper 30 cm of the beach. The depth of evaporation is probably greater in tropical beaches. However, the formation of beachrock below a dry sand cover as thick as 100 cm (Daly, 1924, p 139), would require considerably deeper evaporation than is likely.

On the other hand, if one agrees to the supersaturation of normal sea water, one might wonder why cementation is restricted to the intertidal zone and does not take place wherever water is in contact with nuclei; as for example, below the low tide mark. This perhaps can be understood in terms of the sluggishness of the equilibration between solid and dissolved calcium carbonate, and the inhibition of this equilibration by organic matter. Miller (1952, p. 171) gives six to eight hours as the time required for equilibration in a small, pure, laboratory system. Hindman (1943) has shown that organic matter inhibits equilibration. Taking Hindman's (1943) value of the apparent solubility product of calcium carbonate in sea water, normal surface water in the tropics is several hundred per cent supersaturated. It is possible that even with such supersaturation precipitation on nuclei does not take place. Only in certain beaches and shallow water areas under the effects of high temperature and the film condition of the residual water can it occur at a significant rate.

The acicular form of the aragonite probably is related to several factors. K. Schultze (quoted by, Kubiena, 1938, pp. 182–3) found that the acicular form of crystals precipitated on and in gels is related to rate of evaporation and alkalinity. Descriptions of beachrock, and the author's observations in the Tortugas, indicate that it occurs on

beaches normally protected, but which receive storm wave action. The outbuilding of such beaches during storms produces a deposit which is not disturbed by normal wave action—a necessary condition for lithification.

Even on individual beaches, beachrock is not continuous. This distribution may be explained as a result of the rate at which a portion of the beach drains at low tide. Following Emery and Foster (1948) the rate of beach drainage is proportional to beach slope and permeability. Because storm-built beaches probably have a steeper slope and generally coarser grain size than normal beaches they might be expected to drain more rapidly. Such rapid drainage would allow more time for precipitation of calcium carbonate interstitially. It seems likely that storm-built beaches have pronounced variations in thickness, texture, and structure along their trends. This condition, and the connection between the storm-built beach and the normal beach will affect the rate of drainage at low tide, and thus localize beachrock.

The absence of beachrock on the few beaches of the Florida Keys is due to the thinness of the beach deposits. Pleistocene rock is within a few feet of sea level over most of the area and thus the beach-sands are only a veneer.

The restriction of marine beachrock to the coral seas indicates that temperature is probably the most important factor in its formation. However, high temperatures and evaporation alone are not sufficient for beachrock formation. Emery and Foster (1948, p. 653) reported the common occurrence of a salt crust on Southern California beaches at low tide. As calcium carbonate is an early precipitate from evaporating sea water one would expect beachrock formation in these beaches. Its absence is probably related to restriction of precipitation to the exposed surface which is reworked by wave action at high tide.

As the tidal range at Loggerhead Key is less than a few feet the presence of eight feet of beachrock there (Vaughan, 1914) requires a relative drop of sea level of the order of five feet. A eustatic change of this magnitude is recognized throughout the coral seas.

FOSSIL BEACHROCK

The writer has found no report of pre-Tertiary beachrock in the literature. The recognition of numerous organic reefs in the geologic record suggests that an associated deposit such as beachrock is to be expected. However, there may be some question as to its preservation. As soon as it is exposed on the shore beachrock is very rapidly attacked by intertidal erosion. This erosion is so rapid and thorough that unless new rock is continually formed there would be very little of it on modern shores. Despite this continual destruction it seems likely that under conditions of rapid burial beachrock could be preserved.

The texture, structure, and composition of beachrock offer no obvious criteria for differentiating it from normal beach, bar, or nearshore deposits. The recognition of a brackish or intermittently-exposed facies might be supporting evidence for supposed beachrock.

The most significant criteria are those of calcareous shorelines in the tropics. The appearance of a buried surface such as that shown in figure 3 in an ancient limestone would suggest the term "solution unconformity." A local unconformity of this type with associated slabs or boulders of beach arenite would suggest beachrock. The presence of attached or boring organisms on or below this type of surface might further suggest its formation in the intertidal zone.

CONCLUSIONS

Marine beachrock is produced by an intertidal precipitation of interstitial aragonite under the influence of increased temperatures and the rate and degree of beach drainage. The restriction of beachrock to the coral seas is due to its temperature dependence and the localization of this rock is related to beach stability and structure.

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FIGS. 1–4, JUNE, 1953

CLAY MINERALS IN THE TYPE SECTION OF THE MORRISON FORMATION

W. D. KELLER
University of Missouri

ABSTRACT

The argillaceous rocks in the type section of the Morrison formation as described by Waldschmidt and LeRoy may be subdivided on a basis of their most abundant clay mineral. Kaolinite and illite clay mineral zones coincide with the upper unit subdivisions as proposed by Waldschmidt and LeRoy. A thick montmorillonite-rich zone near the center of the formation transects three of the Waldschmidt and LeRoy units. Kaolinite and illite zones are present in the lower unit.

The clay minerals are used as indicators to interpret tentatively the environments of deposition for different parts of the Morrison.

INTRODUCTION

Geologic formations are regularly subdivided into units or zones which are distinguishable lithologically or mineralogically by the unaided eye in order to serve certain purposes of classification, organized study, or convenience. Argillaceous rocks, however, are not so readily subdivided into clay mineral zones because of the extra instrumental data necessary to differentiate the clay minerals, and therefore few mineral description of thick argillaceous formations have been published. In recent years, however, those rocks have received increasing attention.

Gude (1950), by the use of X-ray diffraction, subdivided the Laramie formation at Golden, Colorado into three stratigraphic intervals based on clay mineral assemblages. The formational boundary was raised 65 feet over a previous location made on other criteria. Van Houten (1948), related clay mineralogy to some of the red-banded characteristics of early Cenozoic deposits in the Rocky Mountain region. Marine shales have been found to be characterized generally by illite (Grim, 1951; Millot, 1949), but the deposition and diagenesis of clay minerals in non-marine rocks have been less widely studied (Grim, 1951). This report on the clay minerals occurring in certain beds of the Morrison formation which is non-marine, will add to the relatively scanty mineralogical information on such non-marine rocks.

FIELD AND LABORATORY WORK

The newer proposed type section of the Morrison (Waldschmidt and LeRoy, 1944) was examined during the summer of 1950 and some 28 specimens of clays and shales, and 9 limestones (Robbins and Keller, 1952), were collected accessory to another research project. This suite of argillaceous rocks, although not ideally complete, is probably numerous enough to give a representative sample of the clay mineralogy of the section. Because individual layers in the Morrison are notably discontinuous and variable both laterally and vertically, the formation is preferably described in terms of zones, or units as was done by Waldschmidt and LeRoy. The specimens at hand are adequate in number to serve this objective. The specimens were numbered to correspond with the beds described by Waldschmidt and LeRoy, plus two others which are designated MU-1 and MU-2.

After collecting, the shale and clay specimens were disaggregated in the laboratory and dispersed in distilled water. Ammonium hydroxide (0.01N) was added to some suspensions to aid in defloccula-

tion. Particles whose diameters were less than 2 microns (equivalent to spheres, Stokes' law) were siphoned off and dried below 50°C. These clay-sized fractions were analyzed by differential thermal and X-ray apparatus.

CRITERIA FOR RECOGNITION AND CLASSIFICATION OF CLAY MINERALS

The kaolin group of clays was identified with X-ray diffraction by a 7A basal spacing in powder diffraction patterns, the illite group by a 10A spacing, and montmorillonite (or montmorillonid, Brindley, et al, 1951) clays by a 17–18A spacing after solvating with ethylene glycol. Kaolinite-containing clays were run a second time after heating to 600°C for 1 hour to destroy the kaolinite basal spacing, to sharpen that of illite, and to aid in the detection of chlorite.

Differential thermograms of Morrison kaolins showed strong endothermic peaks at 575° to 595°C, and exothermic peaks at 945° to 970°C, which correspond to those of reference type kaolinites. Moreover, clays which recorded typical kaolinite interplanar spacing also gave typical DTA reactions, hence correspondence between the data by the two analytical techniques is good.

More variation exists, however, between the X-ray and DTA for the Morrison 2:1 layer clays. Some montmorillonites, as indicated by X-ray, show OH differential endothermic reactions in the 660°–680°C range which may be considered typical, whereas other montmorillonites (by X-ray) show clear cut OH differential endothermic reactions in the range 555°–580°C.

The latter endothermic temperature range is characteristic of ferroan montmorillonite (nontronite) or of illite, rather than of "normal" montmorillonite. Some of the 2:1 clays give endothermic reactions in both the 550°–580°C and 660°–680°C ranges, which may be interpreted as possibly indicating mixtures within the montmorillonite group, mixtures of montmorillonite and illite, or as clays which are transitional in diagenesis or in degradation between montmorillonite and illite. These will be discussed later. The Morrison 2:1 layer clays usually show endothermic reactions also at 140°–160°, and 230°–240°C.

While describing the laboratory procedure which was followed, mention may be made of the effect of mineral orientation on the sensitivity of X-ray diffraction apparatus to record small amounts of clay minerals. During the first part of this study, diffraction films were made of clay samples mounted presumably in random particle orientation on an oscillating wedge holder (General Electric). Differential thermal analyses were run concurrently of the same clays, but these curves showed the presence of small amounts of accessory clay minerals which were not visibly recorded on the X-ray films. This lower sensitivity of X-ray detection is not an uncommon observation, but nevertheless confirmatory evidence by two independent methods strengthens any mineral identification. Later, it was found that if the same clays were mounted with parallel orientation of the platy particles in the conventional sample holder of a Geiger tube, X-ray spectrometer (North American Philips) the threshold of instrumental detection was reduced (sensitivity to small amounts increased) to about the same order as with DTA, and the prior results from DTA were confirmed. The increased sensitivity arises from the tendency of the clay plates to pack parallel to each other and parallel to the surface of the specimen which is exposed to the X-ray beam, as occurs if the clay is pressed into the specimen holder against a removable glass microscope slide during mounting. The 001 spacing record becomes much intensified. The same principle of using an oriented clay specimen was described in 1937 by Bradley, Grim, and Clark for film work, but the preparation of a specimen with preferred orientation is even simpler using the spectrometer-type holder.

Obviously certain disadvantages may accompany the advantages of preferred

orientation of particles in the X-rayed specimen, and some specimens should be run also with random orientation of particles. McCreery (1949) has described a technique for maintaining randomness in the specimen during mounting.

All of the Morrison samples were rerun with X-ray so that both types of mounting were employed.

DESCRIPTION OF SELECTED BEDS IN THE MORRISON TYPE SECTION

Waldschmidt and LeRoy described bed by bed their proposed new type section for the Morrison and grouped the beds into lithologic units as indicated in table 1 below, which is taken from their paper (1949). In order to save space, the full descriptions only of beds sampled for this clay study are reproduced.

CLAY MINERALS IN THE SECTION

The clay minerals which were identified in the less than 2-micron size fractions of the above described argillaceous Morrison beds are listed in table 2. In the first column on the left are listed the clay minerals as determined by X-ray spectrometer. In the third column to the right are mineral determinations by DTA. The letter symbols which are used are explained below the table. The dominant mineral, where more than one is present, is listed first in the series. If montmorillonite was shown typically by both X-ray and DTA it was designated by M; where the X-ray showed the montmorillonite basal spacing (17–18A in glycol) but the DTA showed an OH endothermic reaction below 602°C, it is designated M′. The temperatures of the DTA peak maxima or minima are listed in detail in the remaining columns to the right.

Kaolinite in the sandstone and shale unit.—All of the samples collected from the uppermost sandstone and shale unit, as designated by Waldschmidt and LeRoy, contain kaolinite as the dominant clay mineral. Illite also is present scantily in all of them, except that it occurs in appreciable quantity in LD-112. Confirmation of kaolinite by DTA is without exception. Therefore, the sandstone and shale unit (W. & L.) is characterized by kaolinite, and the clay mineralogy of the argillaceous beds supports the lithologic subdivision as made by Waldschmidt and LeRoy.

Illite in the red shale unit.—The red shale is characterized by illite being dominant and kaolinite subsidiary, with one exception, LD-105, where the order of abundance is reversed. Again, the break in clay mineralogy accords with the division as designated by Waldschmidt and LeRoy.

The X-ray spectrogram of LD-12, taken as a representative pattern of the Morrison illite, solvated in glycol is shown in figure 1. The peak at about 10A is broad, and somewhat more blunted than is shown by type illite. Some kaolinite, quartz, and a slight amount of

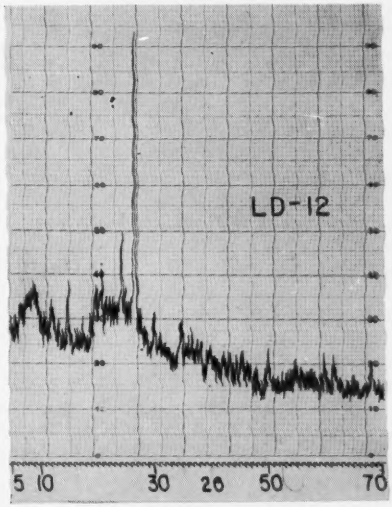

FIG. 1.—X-ray spectrogram of LD-12 solvated in ethylene glycol. Degrees 2θ are indicated at the bottom. The 001 peak of illite is located at 8.8° corresponding to an interplanar spacing 10.1 A. A small kaolinite peak is shown at 12.5°—7.1 A., and the more prominent lines of quartz and calcite are also present. Ni-filtered, CuK_α radiation.

TABLE 1.—*Section of the Morrison formation (modified after Waldschmidt and LeRoy)*

		Feet
	Sandstone and Shale Unit (Top of Morrison)	
LD-1	Shale: slightly sandy, maroon and light lavender intermixed. Lavender shale predominant in top 6 inches, somewhat platy; contains few poorly preserved plant impressions	6
LD-2	Shale: sandy, greenish-gray, faint lavender, pale maroon	6
	Sandstone	11
	Shale	1
	Sandstone	11
	Shale	3
	Sandstone	8
MU-1	Shale: gray, maroon, yellow intermixed; sandy, moderately bedded	3
	Sandstone	7
	Shale	2
LD-112	Shale: dark maroon, sandy, poorly-bedded, blocky fracture, some greenish-gray streaks	3
LD-7	Sandstone: lower 4 feet contains numerous layers of light gray, slightly gritty, hard, compact shale	15.5
	Total	76.5
	Red Shale Unit	
	Shale	1
LD-9	Shale: gray to maroon, sandy, blocky fracture, white, speckled, sand grains medium	4
LD-109	Shale: maroon, sandy blocky fracture	3
	Sandstone	1.5
	Shale	1
	Shale	2
LD-10	Shale: gray, slightly sandy, marine, compact, irregular fracture, slightly speckled, slightly calcareous	7
LD-105	Shale: pale greenish-gray to pinkish, gritty, very hard, compact, massive, uneven to sub-conchoidal fracture	3
	Sandstone	1
LD-12	Shale: light gray to pale buff, mottled, sandy, compact, irregular fracture. Pale maroon in lower foot. Two feet of medium-grained friable sandstone lens in middle part	6
	Sandstone	1
	Shale	0.75
	Sandstone	0.5
	Shale	5.0
	Total	36.75
	Gray Shale and Sandstone	
	Shale	4
LD-13	Shale: gray, sandy, slightly calcareous, massive, hard, compact, uneven fracture	1
	Clay	1
LD-27	Clay: gray, massive, hard, locally slightly calcareous	12
	Sandstone	1
	Shale	2
	Sandstone	0.5
	Clay	2
	Limestone	1.5
	Clay	2
LD-34	Clay: bluish-gray, soft, locally slightly calcareous	5.5
	Sandstone	1.5
LD-39	Clay: gray, soft, slightly calcareous	3
	Limestone	0.5
	Dolomite	2

TABLE 1.—*Continued*

		Feet
LD-43	Clay: gray, massive, uneven fracture, hard	1.6
	*Limestone, LD-44	2.5
	Sandstone	8
	Total	51.5
	Gray Clay and Limestone Unit	
LD-48	Clay: gray, massive, soft, uneven fracture	2
	Limestone	0.5
LD-20	Clay: gray, massive, poorly-bedded	3
	*Limestone, LD-51	1
	Clay	2
	Sandstone	2
	Clay	1.5
	Limestone	0.5
	Clay	2
	Sandstone	2
	Clay	0.5
	Limestone	0.25
	Clay	1
	*Limestone, LD-62	1
	Clay	0.5
	*Limestone, LD-64	2
LD-65	Clay: gray, massive, poorly bedded	3
	Limestone	0.5
LD-67	Clay: gray, massive, poorly bedded	3
	Limestone	0.25
	Clay	3
	Limestone	1
	Clay	1
	*Limestone, LD-73	1
LD-74	Clay: gray, massive	1
	Limestone	1
LD-76	Clay: gray, massive	2
	*Limestone, LD-77	1
LD-80	Clay: gray, massive. Slightly calcareous in lower 4 feet. Ostracod sp.	8
	*Limestone, LD-81	1
	Total	49.75
	Gray and Red Shale Unit	
LD-83	Clay: gray, massive, locally sandy	12
	*Limestone, LD-85	1
	Clay	0.5
	Limestone	1
LD-89	Clay: gray, massive, poorly bedded, *Aclistochara* in lower 5 feet	8
	*Limestone, LD-91	1
LD-92	Clay: gray, massive, sandy to silty, *Aclistochara* in top foot	8
MU-2	Shale: red to gray, color unevenly distributed, mottled appearance	24
	Total	55.6
	Basal Sandstone Unit	
	Sandstone	7
	Total thickness of formation	277 feet

* Asterisks indicate limestones whose non-carbonate minerals are described.

calcite are also indicated. The spectrogram is fairly typical of clay-size fractions initially separated from illitic shale.

Kaolinite in the gray shale and sandstone unit.—The upper 14, perhaps 18, feet of the gray shale and sandstone unit

TABLE 2.—*Clay minerals identified in the Morrison samples*

X-ray determination	Bed and sample number	DTA determination	N	N	N	N	N	N	N	X
		Sandstone and Shale Unit (W. & L.)								
K, Iw	LD-1	K					575			960
K, Iw	LD-2	K					595			960
K, Iw	MU-1	K					575			945
K and I	LD-112	K and I	150	280			580		910	960
K, Iw	LD-7	K, I (?)		220			580			970
		Red Shale Unit (W. & L.), Clay Mineral Change								
I, Kw	LD-9	I	150	230		560			850	950
I, Kw	LD-109	I	150	230	435	545				900
I, K (?)	LD-10	I	150	280	445	525				880
K, I	LD-105	K, I	140				580			960
I, Kw	LD-12	I	150	230		550				925
		Gray Shale and Sandstone Unit (W. & L.), Clay Mineral Change								
K, Iw	LD-13	K, Iw	150			550				950
K, I	LD-27	I, K (?)	145	225		540			870	980
		Clay Mineral Change								
M, Iw, Kw	LD-34	M′	155	225		560			900	990
M, Iw, Kw	LD-39	M′K (?)	150	240		570			860	950
M, I, K	LD-43	M, I	145	235		560		660	850	935
		Gray Clay and Limestone Unit (W. & L.)								
M, Iw, Kw	LD-48	M′, K (?)	150	230		570			900	970
M, I, K	LD-20	M, I	160	240		580		680		970
M	LD-65	M, I (?)	175	245		555		670		975
M, Kw	LD-67	M	160	240				660	860	980
M, Iw	LD-74	M, I (?)	160	240		560		680	800	900
M	LD-76	M, I (?)	160	240		580		680	850	960
M, I (?) K (?)	LD-80	M, I (?)	160	240		580		700	890	
		Gray and Red Shale Unit (W. & L.)								
M	LD-83	M′	160	220		560			880	
		Clay Mineral Change								
K, M (?)	LD-89	M′, K (?)		220		570				
K, I	LD-92	K, I	140	195			570		880	960
I, K (?)	MU-2	I	160	220		550				850

Symbols:
I—Illite
M—Montmorillonite indicated by both X-ray and DTA
M′—Montmorillonite by X-ray, OH endothermic reaction below 600°C
K—Kaolinite
N—Endothermic reaction peak
X—Exothermic reaction peak
w—weak, scanty, or minor in amount
(?)—Such slight instrumental response as to be uncertain of mineral occurrence.

are characterized by kaolinite as the dominant clay mineral, with illite present as a minor constituent. Below this zone occurs a pronounced break in the clay mineralogy.

Montmorillonite zone transecting three lithologic units.—Montmorillonite, and probable montmorillonite-derived, clay minerals characterize the next descending 100 feet of the type Morrison section. This montmorillonite-dominant zone extends downward through the remainder

of the gray shale and sandstone unit, throughout all of the underlying gray clay and limestone unit, and continues below through the upper 12 feet (LD-83) of the lowest argillaceous group, the gray and red shale unit. In the upper part of the zone, illite and even kaolinite occur in small amounts with the montmorillonite, but they diminish to *nil* in much of the lower part of the zone.

Specimen LD-67 illustrates the more typical variety of montmorillonite. Its DTA endothermic peaks occur at 160°, 240°, 660°, and 860°C, and an exothermic peak occurs at 980°C. Its X-ray spectrogram shows a 17A-18A basal spacing when solvated with glycol, and other subsidiary peaks, figure 2.

Other montmorillonite (by X-ray) specimens which show OH endothermal reactions below 600°C are designated by M′ in table 2. Specimen LD-39 was selected as a representative of this group and was studied in most detail. An X-ray spectrogram of it solvated in glycol, packed so as to develop parallel orientation of clay plates, is shown in figure 3, and one of its dry powder mounted in random orientation is shown in figure 4.

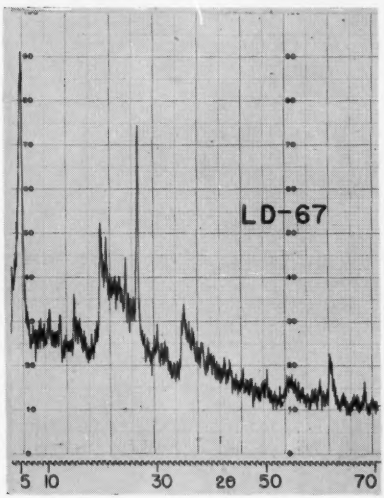

FIG. 2.—X-ray spectrogram of LD-67 solvated in ethylene glycol. Degrees 2θ indicated at the bottom. The 001 peak of montmorillonite is located at 5.2°, corresponding to an interplanar spacing of 17 A. A slight amount of kaolinite probably is indicated at 12.9°. Ni-filtered, CuK_α radiation.

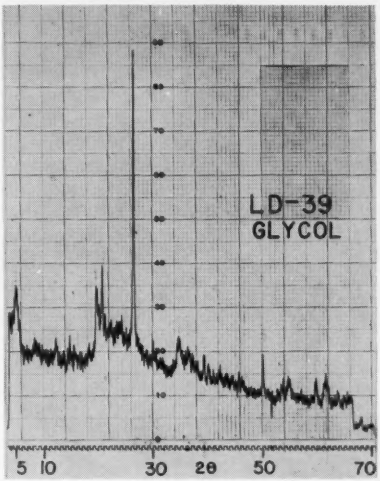

FIG. 3.—X-ray spectrogram of LD-39 solvated in ethylene glycol. The 001 peak at 5.2°—17A is less intense than the 001 peak of dry, unsolvated LD-39, shown in figure 4. The 001 peak of typical, pure montmorillonite is significantly intensified and sharpened when the clay is solvated in ethylene glycol. Apparently LD-39 is not pure montmorillonite.

Note the weaker diffraction in the range of 19°–27° of LD-39 in figures 3 and 4, than of LD-67 in figure 2. These anomalous characteristics of LD-39 are similar to those exhibited by a mixed layer, 2:1 clay, and it is thought that LD-39 is therefore a mixed-layer clay. Ni-filtered, CuK_α radiation.

Reference to figure 3 shows a montmorillonite peak at about 17A, small peaks at 10.4A (illite) and 7.14A (kaolinite), and peaks corresponding to quartz. The spectrogram of the dry (not solvated with glycol) powder in random orientation, figure 4, shows a more pronounced montmorillonite peak at about 11.8A than is shown where the clay is solvated (figure 3), and accessory kaolinite and quartz. The illite peak (?)

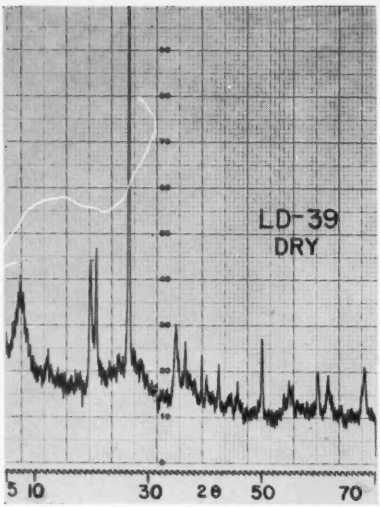

FIG. 4.—X-ray spectrogram of dry, unsolvated LD-39. Degrees 2θ are indicated at the bottom. Note the prominent peak of unexpanded montmorillonite clay at 7.4°, corresponding to 12A. The pattern between 19° and 27° is not that of typical montmorillonite. Ni-filtered, CuK_α radiation.

may be masked by the montmorillonite. For unsolvated montmorillonite to register a stronger 001 peak than where it has been solvated with glycol, is inconsistent with the instrumental response from monomineralic montmorillonite. It occurs instead in clay of the mixed-layer type, as in the API Reference Clay No. 31, Mixed Layer Aggregate from Cameron, Arizona.

In mixed layer 2:1 clay the regularity of lattice repitition in the dry clay may be interrupted by a somewhat randomized or only partially repetitive 16A–18A spacings swelled by glycol solvation, and presumably thereby the intensity of the 001 spacing be reduced. The mechanism giving rise to interrupted repetition of 001 spacing would be similar, except in reverse order, as was described and illustrated by Jackson et al. (1952) in the "depotassication" of a layer silicate. With monomineralic montmorillonite, or with illite, the effect of solvating and packing is to enhance the intensity of the 001 spacing.

Chemical analyses of the clay from Cameron, Arizona, and LD-39 are given in table 3.

TABLE 3.—*Chemical analyses of clays*

	A	B
SiO_2	54.37	56.90
Al_2O_3	17.79	16.72
Fe_2O_3	7.14	3.76
FeO	0.27	0.17
MgO	1.20	1.09
CaO	1.83	2.68
Na_2O	0.26	0.11
K_2O	2.49	3.06
P_2O_5		0.28
H_2O+	6.79	10.24
H_2O-	6.31	4.88
TiO_2	0.56	0.05
CO_2	0.32	n.d.
C	0.17	n.d.
Total	99.50	99.94

A. Analysis of the API Reference Clay No. 31, mixed Layer Aggregate, Cameron, Ariz. From page 56, API Reference Clay Minerals, Preliminary Report No. 7.
B. LD-39. Analysis by Bruce Williams Laboratories, Joplin, Mo.

Referring to the chemical analysis, B, of LD-39, the low contents of Fe_2O_3 and FeO preclude the clay from being nontronite, which is to be considered in terms of an OH endothermic peak below 600°C. The high content of K_2O falls within the range of illite, but the X-ray pattern rules out illite as the dominant mineral. Hence, LD-39 is considered also on the basis of composition to be a mixed-layer type of 2:1 clay. The other Morrison clays designated M′ are probably also mixed layer clays.

From the viewpoint of stratigraphic petrology, the variation in montmorillonitic clays is presumably less significant than the unbroken sequence of montmorillonite-dominated clay down into the gray and red shale unit. If different clay minerals were as easily differentiated by megascopic tests as are dolomite and calcite, for example, it is probable that Waldschmidt and LeRoy would have

entertained an alternative unitization of the Morrison such as is indicated by the clay mineral group changes shown in table 2.

Clay mineral zones become important as stratigraphic units insofar as they can be recognized as indicators of the kind of sedimentational environments which prevailed. Progress in this direction will be discussed later.

Kaolinite and illite in the lower Morrison.—Kaolinite predominates in the middle part, and illite in the lowest shale of the gray and red shale unit. *Aclistochara* were reported in these kaolinite-rich beds.

SUMMARY

Changes in the clay mineralogy of the argillaceous beds support the subdivision of the Morrison formation at the new type section into the three uppermost units proposed by Waldschmidt and LeRoy. A single montmorillonite-dominant zone carries downward across their next three lithologic units. Waldschmidt and LeRoy however qualified the divisions of those latter units by noting that the probability of a formational hiatus between the gray shale and sandstone unit and the gray clay and limestone unit "can not be evaluated at this writing; and that the base of the gray clay and limestone unit is arbitrarily placed at the highest observed occurrence of *Aclistochara*."

Kaolinite and illite predominate in the middle and lower portions, respectively, of the gray and red shale unit.

ENVIRONMENTAL INTERPRETATION FROM CLAY MINERALS

Clay minerals are notably sensitive in a qualitative way to their physical and chemical environments during deposition and diagenesis, and therefore the clay mineral which results from sedimentation becomes an indicator of the nature of those environments (Ross, 1943; Ross and Hendricks, 1945; Keller, 1946; Millot, 1949; Grim, Dietz, and Bradley, 1949; Keller and Ting, 1950; Grim, 1951; Keller, 1952). The extent to which clay minerals and argillaceous rocks can be used as indicators of the environments occurring in the geologic past depends largely, however, upon the ability of the observer to make a judicious interpretation of the mineral evidence before him.

Some progress has been made in this direction but the advance is probably not far beyond the beginning stage. A logical approach is to proceed from the known to the unknown, i.e., from a study of present day sediments to those of the past, as has been done in papers by Millot (1948) and Grim, Dietz and Bradley (1949). However, some benefit may arise also, by examining ancient sediments such as the Morrison, and observing the wide gaps and inadequacies in the interpretation of their geologic history which need to be filled in. These will become apparent in the following discussion.

The uppermost unit, the sandstone and shale unit, of the Morrison at the type section contains kaolinite as the dominant clay mineral. Limestone and calcareous marl are notably lacking, a condition which is consistent with the usual occurrence of kaolinite. In environments both of weathering (Ross, 1943) and of deposition (Millot, 1949), kaolinite genesis is favored by an abundance of circulating "renewed," "aggressive" fresh waters which are relatively low in metal cations but have a H^+ concentration preferably high enough to be acid, and are oxidizing in redox potential. This environment is consistent with the nonmarine origin which is generally accepted for the Morrison.

Below the kaolinite-bearing member lies the red shale unit which is characterized by dominant illite, the common mineral in marine shales. The structure of the clay in this unit is regularly "blocky, compact, massive" (Waldschmidt and LeRoy), rather than being thinly laminated which is more typical of many marine shales. The dominant illite content in this unit is a significant mineral change from the kaolinite above,

and merits considered attention. Because illite is the characteristic clay in a marine shale (Millot, 1949; Grim, Dietz, and Bradley, 1949; Grim, 1951), its presence at first thought appears to be incompatible with a non-marine origin for the Morrison.

However, a broader approach to the genesis of the clays emphasizes that indicator clays are formed as a response to a set of physico-chemical conditions rather than in an artificially classified environment, such as *marine*. Fundamentally, the formation of illite is favored in a body of water where the solutions are not frequently renewed or freshened, where the pH is alkaline, where cations (metal) are very abundant and potassium predominates, and the solubility and stability of SiO_2 and Al_2O_3 in the aqueous solution gives rise to a mineral whose SiO_2/Al_2O_3 ratio exceeds 3 (Millot, p. 290).

These chemical conditions usually prevail under the normal saline (NaCl) marine environment, and therefore illite characterizes marine shale. However, if the metal cations, especially potassium, are abundant, and the water is alkaline, it may still be "fresh" as regards low NaCl content, and therefore predispose aluminum silicate compounds toward the illite group. Underclays in which the clay mineral is almost solely illite probably originated in fresh to brackish water where calcium carbonate was also being deposited (Grim and Allen, 1938; Grim and Bradley, 1939).

Hence the transition between the illitic lower red shale unit, upward into the kaolinitic sandstone and shale unit probably followed increased "aggressiveness" (Millot) and freshening of the water, more frequent renewal by circulation, less K^+ and more H^+, or alternatively a different source of clay, and rapid deposition and covering of the kaolinite (Grim, Dietz, and Bradley).

As regards lamination in shale, or massiveness in various other claystones, the writer believes that laminations in shale arise mainly from the parallel packing of clay particles, which is favored by the clay having been flocculated into tiny platelets before it has settled to the bottom, and/or that deposition took place slowly enough and in a suspension sufficiently dilute that no considerable interference took place between adjacent platelets during deposition (Keller, 1936, 1946). On the other hand, randomly oriented clay mineral crystals as in some non-laminated claystones, may have crystallized or grown within a clay gel, or the clay may have been dumped into the basin rapidly, or by a high-density (turbidity-type) flow, in which either prior turbulent motion or high viscosity of the mud accounted for a non-parallel packing of clay particles.

Descending from the red shale unit into the subjacent gray shale and sandstone unit, kaolinite is the dominant clay mineral over an interval of 18 feet. Here, as in the top unit (the sandstone and shale unit) kaolinite is accompanied by an abundance of sand. Does this repeated association of quartz and kaolinite in the first and third units from the top mean similar provenance and/or similar environments of deposition for those units? Does the intervening illitic red shale unit represent an interruption in the kaolinite-sand period of deposition, or is it an alternation in environmental conditions, something like a recession of this locality from a prior, closer-to-high land position? Does the abundance of sand associated with the kaolinite mean a source of sediment at a relatively higher elevation and faster flowing streams or more rapid deposition without time enough to wash and sort clay and sand size particles; or a different terrain, climate, and pronounced leaching of a permeable sandy soil to furnish quartz and kaolinite, in contrast to the kind of source and deposition of the intervening illite-bearing red shale unit? It is logical to return affirmative answers to these questions in the absence of evidence to the contrary, but it should be recognized that although the postulates are logical and probable, no one of them

is necessarily the true explanation for the observations. Geologic conclusions must unfortunately arise in many cases from only permissive evidence because the physical conditions which surround the problem limit significantly the amount or the incisiveness of the data which can be obtained. An affirmative answer to the above questions should be taken as only tentative; a detailed mineralogical study of other Morrison sections scattered over a wider area should provide data upon which conclusive answers be returned to the questions.

Below the kaolinite-dominated zone lies about 100 feet of gray shale, gray clay, sandstone, and limestone and dolomite, in which the clay rocks are dominated by montmorillonite and mixed layer 2:1 clay minerals. The clay beds are commonly calcareous to marly, and limestone and dolomite layers, usually argillaceous to sandy, are abundantly interspersed among the clay and sand beds. Dinosaur remains, fossil spicules, and *Aclistochara* in one limestone layer occur in this part of the section. Examination of the sand-size particles from a few random clay samples in this montmorillonite zone show occasional feldspar, mica, high index, probable heavy minerals, very sparse isotropic (glass?) fragments, and many aggregates of microcrystalline alteration products. These sands are thought to represent volcanic dust particles well along in alteration, and confirm the observations of others who have long considered portions of the Morrison to have been derived in a large part from fragmental volcanic material.

Montmorillonite is a common alteration clay from volcanic dust of salic to intermediate composition in an environment where hydrolysis is effective, but removal of cations is not drastic. Such an environment includes that which would apply to the weathering of volcanic dust as it is shifted about on alluvial fans, flood plains, wind-deposited mantles, and in poorly drained marshes or interior lakes, which appear from field observations to have been sites for deposition of Morrison sediments.

The montmorillonite-illite mixed-layer clays are thought to be transition products between montmorillonite and illite via the route of potash fixation, here only partially completed (Volk, 1933, 1934; Wear and White, 1951). Potash fixation was likely augmented because of a probable abundance of soluble silica which would have been released from the hydrolyzing volcanic glass (Mortland and Geiseking, 1951). The potash-bearing clays of the Morrison are thought to be products of constructive diagenesis toward illite, rather than illites which have been degraded (Grim) or "depotassicated" (Jackson, et al. 1952).

The montmorillonite dominated zone was developed because of the abundance of volcanic dust and the weathering environment which surrounded it during deposition.

Illite is abundant, usually the dominant clay mineral, in limestones and dolomites which were collected from this part of the Morrison section, but kaolinite and montmorillonite are found in two of them as shown below (Robbins and Keller, 1952).

LD-44	Kaolinite strongest in X-ray, illite strongest in DTA
LD-51	Illite
LD-62	Montmorillonite, illite, kaolinite (?)
LD-64	Illite, kaolinite, montmorillonite all weak but about equal intensity in X-ray patterns
LD-73	Illite
LD-77	Illite, kaolinite (both weak)
LD-81	Illite, diffuse kaolinite (?) band

Apparently the development of illite was facilitated by submergence under water which was alkaline and saturated with Ca (and Mg in dolomite basins) while limestones were being deposited.

Below the montmorillonite zone, in the gray and red shale unit, kaolinite is dominant in about 16 feet of massive clay which contains *Aclistochara* fossils. Limestone LD-85, taken from the top of this kaolinite zone, contains clay which showed kaolinite strongest in X-

ray and illite strong in DTA. An influx of kaolinite to the site of deposition at a rate greater than that of alteration to illite (assuming that illite would be most stable, at least in the lake where limestone was deposited), would account for the presence of a kaolinite zone. Limestone LD-91 from within the clay contains illite and questionable kaolinite. It is logical, permissively, that the clay became illite in LD-91 under the environment of limestone deposition.

Aclistochara are associated with both kaolinite and montmorillonite clays (also mixed-layer 2:1 clays) but the range in environments in which present-day forms live is sufficiently broad to include the genetic environments of both types of clay minerals. Wood, a foremost student of modern chara, wrote (1952); "Chara and Characeae in general are predominantly restricted to fresh water. Brackish water forms in North America are restricted to the genus *Chara*, with *Nitella flexilis* invading in some cases up to 4 0/00 NaCl which is the maximum concentration tolerated to the best of my knowledge. In fresh water, chara ranges in tolerance from pH of 9.2 to pH 5.7. I am sure these are not the extremes to which various charas can reach, for in a study of Haunk's Pond, Middle Bass Island, Lake Erie, in 1940, the pH ran above this figure during the day, almost to pH 11, as I recall [the exact figure is uncertain]; and this pond supported a lush growth of several species, being dominated by *Chara zeylanica*. Alkaline (saline, I presume) pools in Utah, were found [by Wood] to contain *Chara evoluta*, but no figures as to pH or salinity were obtained.

"Charas seem to inhabit most any suitable water system in which extremes are not existent. Off hand, the only ones which occur to me in which I would not expect to find some charas would be Sphagnum bogs. . . . , extremely swift rivers and brooks, rivers and still bodies in which turbidity is extreme, stagnant water in which oxygen concentration regularly is very deficient or where blue green algae have taken over. They live at depths ranging from lying in seepage water and those exposed largely to air, such as in fields and seepage ditches I saw at Drumnadrockit, Scotland, and fields at Taunton, Surrey, England; to 40 foot depths, or to 60 feet in depth as reported by Allen in the St. Clair River."

It seems probable that the range which chara tolerate easily encompasses the depositional environments of the Morrison formation.

The lowermost argillaceous member, MU-2, a shale 24 feet thick in the Waldschmidt and LeRoy section, contains dominantly illite, which is typical of shales.

SUMMARY

The zones of dominant clay minerals in the argillaceous members of the Morrison formation at the type section have the following approximate thicknesses:

Kaolinite	77 feet, top
Illite	37 feet
Kaolinite	18 feet
Montmorillonite	95 feet
Kaolinite	16 feet
Illite	24 feet, base

The thick montmorillonite zone near the center is interpreted as originating from altered volcanic dust accompanied by partial potassium fixation. Illite zones are interpreted as originating under conditions of alkaline pH and abundant cations, whereas kaolinite members have been deposited rapidly, or where cations were either removed or at least were not present in high concentration. These interpretations follow Millot's work.

Detailed mineralogical studies of other Morrison sections are needed to test the degree of probability of the tentative environmental interpretations of this section made on permissive evidence.

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FIG. 1, JUNE, 1953

WEATHERABILITY OF ZIRCON

DOROTHY CARROLL
Bryn Mawr College, Bryn Mawr, Pennsylvania[1]

ABSTRACT

Observations made on numerous grains of zircon from the heavy residues of soils from different areas indicate that zircon becomes corroded under certain conditions, particularly those of strong alkaline leaching associated with the development of a lateritic type of profile. Zircon crystals have a number of weak points, such as inclusions and zoning, which enable corrosion to set in; some of these are illustrated by photomicrographs. It is suggested, following information obtained from spring and ground-water chemical analyses, that Ca-bicarbonate and Na-bicarbonate rich waters may be able to dissolve zircon. Several experiments in the artificial weathering of zircon are described; one of these used a synthetic cation exchange resin. The results, obtained over a short period of time only, were negligible.

INTRODUCTION

The most stable of the heavy minerals found in sedimentary rocks, unconsolidated sediments, and in soils is, by consensus of opinion (Smithson, 1950), zircon, and because of this stability and its almost universal presence in heavy mineral assemblages, it has been made use of by Haseman and Marshall (1945) in studies of soil development. In order to assess the value of zircon for this purpose, it is advisable to inquire more closely into its stability in soils and in other environments, and to distinguish clearly between loss by abrasion and loss by weathering processes.

Many minerals originally considered to be more or less unalterable, are now found, when examined closely, to lose substance and become corroded through the long-continued action of solutions with which they come in contact (Edelman and Doeglas, 1934). In his Anniversary Address to the Geological Society of London in 1941, P. G. H. Boswell stressed the importance of change and loss of mineral grains in sediments by solution as one of the most important to be considered by workers in the field of mineralogical aspects of sedimentary rocks, particularly in regard to the interpretation of provenance and past history of such rocks.

This paper records observations which have been made on the appearance of zircon grains, and their stability, in the heavy residues of soils; it also reports an attempt at the artificial weathering of zircon carried out over a short period of time.

PHYSICAL FEATURES OF ZIRCON

Chemically zircon is an orthosilicate, $ZrSiO_4$, in which part of the Zr is often replaced by Hf, Yt, Ce, etc. The size of the Zr^{4+} ion (0.87 A) leads to a slight instability of the structure, so that when radioactive elements are present a metamict state eventually results because of the weakening of the bond between Zr and the SiO_4 tetrahedra.

Three distinct types of zircon have been recognized by mineralogosts: (1) Normal; (2) Hyacinth or purple zircon; (3) Malacon. A study of the distribution of these types in the Lake Superior area pre-cambrian rocks was made by Tyler *et al* (1940), and Morgan and Auer (1941) have given a useful outline of their optical properties:

(1) Normal type—high refringence and birefringence (<0.05).
(2) Hyacinth type—slightly lower refringence and birefringence (0.017–0.043)
(3) Malacon type—refringence much lower and birefringence ± 0.008.

[1] Now at the United States Geological Survey, Washington, D. C.

The radioactivity increases from (1) to (3). In addition to reducing the refractive index and birefringence, the change to the metamict state increases the volume thereby reducing the specific gravity; furthermore, the material can now take in water which gives it a clouded altered appearance. In reporting zircon *from soils and sediments the type should* be stated, for the presence of malacon could be mistaken for genuine weathered normal type zircon.

Zircon has a hardness of 7.5, similar to that of staurolite and tourmaline. In abrasion tests reported by Freise (1931) it has a relative *"transportwiderstand" of* 265, compared with quartz, 245; rutile, 455; garnet, 375; epidote, 320; *staurolite,* 420; tourmaline, 650–950; and apatite, 275. The general smallness of grain size probably accounts for its persistence along with much more abrasive-resistant minerals.

Zircon has no cleavage, but there is a brittleness which causes the breaking into irregular fragments with conchoidal surfaces. It has been observed that there is often a weakness in the structure at the junction of the prism and pyramid faces.

Zircon crystallizes in the tetragonal system, the principal forms being (100), (110), and (111). In addition it exhibits a fascinating number of varietal forms based on the variation of prism and pyramidal faces, on zoning, and on inclusions and their type and arrangement. Brammall's description (1928) of the varietal features and their use is still the most important paper written on this aspect of zircon.

OBSERVATIONS ON CORROSION OF ZIRCON

Observation has shown that zircon grains can be reduced in size by the ordinary processes of weathering, but the actual processes which produce these results are not yet clear. However, there are a number of ways in which zircon can be reduced in size and eventually disappear.

The type of zircon.—It has already been mentioned that there are three types of zircon and innumerable crystallographic varieties. Corroded grains appear to fall into the following groups, considering only Type (1), Normal zircon:

A. Grains with numerous inclusions are less stable than those with very few. *The inclusions may be,* as summarized by Brammall (1928):

(a) gas-filled cavities.
(b) stringers, clots, and cloud-like suspensions of opaque dust;
(c) crystalline inclusions of various minerals such as monazite, hematite, quartz, apatite, etc.

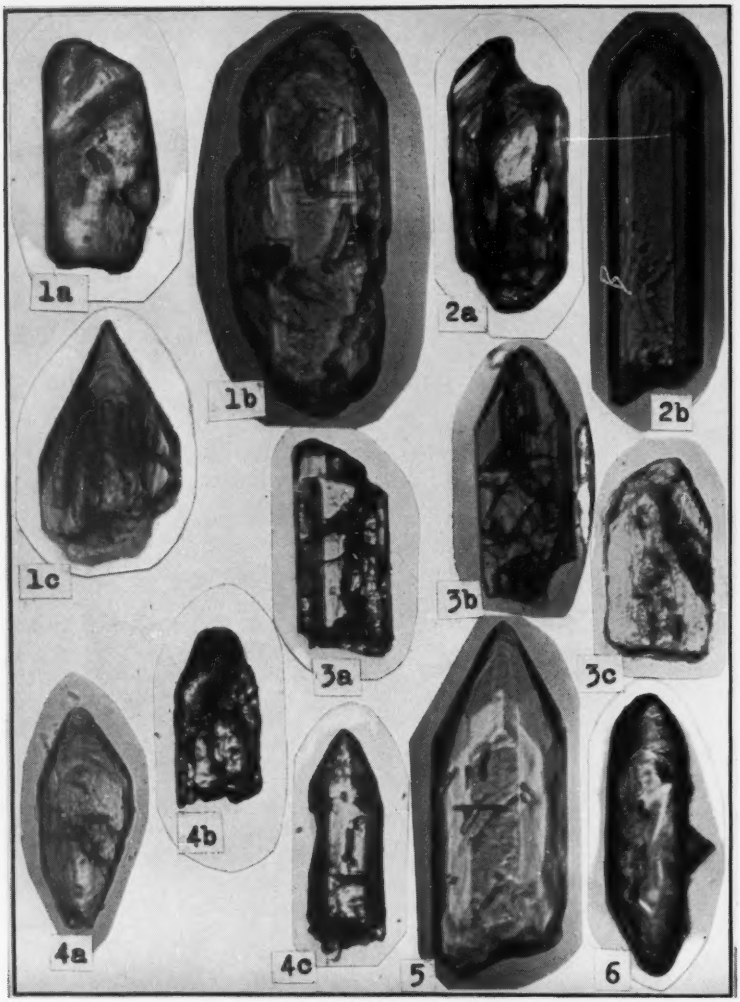
B. Grains which are zoned are less stable than unzoned grains. Zoning appears to be more frequent in the hyacinth than in the normal type zircon.

The action of water containing various solvents appears to be:

(i) Penetration into cavities in the mineral; this is done by the removal of surface material of the grains, particularly when the cavities are only covered by a very thin layer of mineral, i.e., when the cavities are very near the surface. This is illustrated in figure 1.

(ii) Penetration of solutions along zoning lines. The zoning appears to be the result of growth similar to that which produces phantom quartz crystals, and the zones are often lined by a deposit of fine dust-like particles which may be hematite, magnetite, etc. The penetration of solutions is probably assisted by these minute particles of foreign minerals. Zoning is due to lamination, and is one of the results of the introduction of material of slightly different chemical composition, including the variation of radioactive elements, during the growth of the crystal. These slight differences in chemical composition may facilitate the removal of mineral substance (fig. 2).

(iii) Zircon is a brittle mineral, which though hard, may crack and fracture.



LEGEND FOR FIGURES 1 TO 6

FIG. 1.—Zircon grains showing cavities and the penetration of solutions. 1a. Rounded grain showing open cavities and the start of further corrosion. Residue from soil, Bullsbrook, W. Australia. ×190 (in refractive index of 1.54). 1b. Euhedral grain, alluvial, Colorado. ×185 (in R.I. 1.72). 1c. Pyramidal end of euhedral grain with many cavities. Residue from soil near Bel Air, Maryland. ×140 (in R.I. 1.72).

FIG. 2.—Zircon grains showing corrosion along zones. 2a. Grain showing a late stage in penetration. Residue from soil, Lake Grace, W. Australia. ×150 (in R.I. 1.54). 2b. Early stage of corrosion, zones just visible. Alluvial, Colorado. ×185 (in R.I. 1.72).

Cracks on the surfaces of faces allow the penetration of solutions, and this eventually leads to a removal of material. Hutton (1950, pp. 690 and 696) has described microfissuring in hyacinth grains, and states that this fissuring is caused by increase of volume due to transformation of the mineral to the metamict state by radioactive elements. The grains figured here, however, show cracks in the normal type of zircon (fig. 3).

(iv) The broken ends of prisms appear to be much more susceptible to corrosion than the surfaces of prism and pyramidal faces, and where the latter remain intact, corrosion is at a minimum. The broken ends of prisms when corroded show pointed remnants of zircon (fig. 4) very similar to the remnants of amphibole, pyroxene, epidote, and staurolite which are of quite common occurrence (Edelman and Doeglas, 1934). The etching of garnet is confined to the surfaces and appears to be somewhat different, although the cause is probably similar. There seems to be a fundamental difference in the atomic structure at the ends of the c crystallographic axis when compared with the ends of the a and b axes. In tourmaline it has been observed that it is the antilogous end which is most receptive to the deposition of new material as secondary growth (Alty, 1933), although outgrowths also occur elsewhere (Krynine, 1946).

In many minerals characteristic etch-patterns occur on crystal faces, and appear alike on similar faces, e.g. in quartz, where such patterns are used to determine compound twining. Difference in etch-patterns may cause differences in corrosion of faces in zircon crystals, and an indication of this is shown in figure 5. Long-continued etching by solutions will result in the destruction of the original crystal faces, and their surfaces will be removed.

(v) There is the, as yet, unexplained presence of secondary outgrowths of zircon as described by Trueman (1912), Butterfield (1936), Smithson (1937), Tyler *et al.* (1940), Bond (1948), and Hutton (1950). It would appear that some zirconium was present in the rock and soil solutions, together with silica, and that stagnant conditions occurred that were favorable for the formation of outgrowths on the zircon grains already there. Krynine (1946) has described outgrowths on tourmaline; the titaniferous minerals will crystallize from solutions provided the conditions are suitable; and albite occurs authigenically. An outgrowth on zircon is illustrated in figure 6; it is from a heavy residue from the Jurassic of Yorkshire, England, kindly given me by Dr. F. Smithson.

MECHANISM OF CORROSION OF ZIRCON

In weathering processes with normal and hyacinth types of zircon the grains

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LEGEND FOR FIGURES 1 TO 6 (continued)

FIG. 3.—Zircon grains with surface cracks. 3a. Broken prismatic grain with cracks filled with clay material (in R.I. 1.72). 3b. Grain with surface cracks and inclusions (in R.I. 1.84). 3c. Badly corroded grain, the corrosion having started from cracks (in R.I. 1.84). All from soil residue, Bel Air, Md. ×140.

FIG. 4.—Zircon grains showing corrosion at broken end of prism. 4a. Euhedral grain with small cavity at position of c axis. 4b. Much corroded grain with irregular pointed remnants. 4c. Similarly corroded grain with possible small outgrowths at end of c axis. All from soil residue, Bel Air, Md. × 140 (in R.I. 1.72).

FIG. 5.—Euhedral grain showing differential removal of surface layers and development of different appearance of faces. Note inclusions; and pock-marks on (110). Alluvial, Colorado. X185 (in R.I. 1.72).

FIG. 6.—Rounded zircon grain with pyramidal outgrowth. Jurassic, Yorkshire, England. Courtesy Dr. F. Smithson. ×200 (in R.I. 1.54).

are reduced by the slow corrosion of the mineral substance which is removed in solution; there is no alteration of the zircon itself whereby an alteration product occurs. The process is apparently similar in action to that which causes the solution of garnet, staurolite, epidote and zoisite, amphiboles, and pyroxenes, but, as is well known, in the latter two minerals there is also the actual change of the mineral, wholly or in part, and in stages, into a chlorite-like mineral. In soils rich in ferromagnesian minerals, both types of alteration can occur together.

It would seem that the zircon reported as altered (e.g. by Goldich, 1938), with cloudy appearance, may be all the malacon type, in which the alteration is caused by radioactivity, and not by weathering. When more observations have been made, it may be found that the alteration by radioactivity has left the remaining mineral substance more in a condition which can be altered by solutions such as occur in the chloritization of amphibole and pyroxene.

Recently Rankama and Sahama (1950) have drawn attention to the fact that although zircon, as the principal repository of zirconium in the lithosphere, appears to be excessively stable, yet there is a zirconium cycle in Nature. The work of Strock (1941) at Saratoga Springs, N.Y. gives evidence of one way in which zirconium may be transported in solution and redeposited.

The mineral waters of Saratoga Springs are remarkably rich in zirconium which Strock (1941) has determined spectrochemically as 35 mg per liter. The source of this Zr is considered to be zircon as there is no other known source for it, nor are any deposits of zirconiferous minerals known in the area from which the spring waters come. The Saratoga waters are strongly carbonated, highly saline, and calcium-rich, and their action has been to dissolve Zr from zircon in the terrain which they drain. Strock further records (1941, p. 868) that "preliminary experiments have confirmed the ability of Ca-bicarbonate solutions to dissolve considerable amounts of Zr from precipitated ZrO_2 and from zircon and eudialyte." Without going into details here, Strock considers that Zr is carried in the Saratoga waters as a stable dicarbonatozirconylate ion, $Ca[ZrO(CO_3)_2]$, in the presence of excess CO_2. This is stable at the acidity of the water (pH 5.5) and Zr is transported in this form; when the water becomes alkaline the Zr is precipitated.

That Strock's ideas may be correct with regard to the solution of zircon by Ca-bicarbonate solutions is suggested by the work of Foster (1950) on the waters of the Atlantic and Gulf Coastal Plains. She finds that, experimentally, the presence of carbonaceous matter such as lignite and soil organic matter, is sufficient to generate CO_2 in sediments, and when this is "dissolved in water enables it to take into solution more calcium carbonate. If base exchange materials are also present to replace Ca with Na, a still greater amount of bicarbonate can be held in solution" (1950, p. 33).

This process, aided by the cation-exchange capacity of the clay-minerals, is probably quite important in soils and may possibly account in part for the corrosion of many minerals, including zircon. It should be remembered that small grains of zircon in a clayey soil may be embedded in aggregates of clay minerals, and that they are passively subjected to any chemical action that may take place in the processes of cation-exchange during and after soil formation. The writer has recently seen zircon grains completely enclosed in clay which penetrated into cracks in the mineral; in one case a broken zircon crystal was held more or less in its natural position by clay. Graham (1941) has explained the mechanism whereby such minerals as feldspars, amphiboles, and pyroxenes, are altered by the presence of acid clay, and such a process, while not achieving the same results with zircon as with more weatherable minerals, must have some effect on the zircon. The results of such corrosion are recorded in the

individual grains.

From another angle, Maurice (1949) found that zircon has a wide range of stability in acid solutions, but it is restricted in alkaline solutions (experiments were done at a temperature of 400° C. and under a pressure of 900 atmospheres). Zircon was synthesized at 400° C. in aqueous solutions. After a series of experiments, it was found that if ZrO_2 and SiO_2 are in the presence of each other in dilute acid solutions, the silicate will form and will be stable. In alkaline solutions it was not possible to obtain a pure crop of zircon crystals. It was found that the maximum alkalinity at which quartz and zircon are stable together does not exceed 0.07 N Na_2O. Zircon is readily crystallized in acid solutions, but its field of stability in alkaline solutions is very restricted. These observations were all made at high temperatures and pressures, but it is possible that the same results may be produced under normal temperatures in rocks when the time is unlimited.

In 1948 Fredrickson made some interesting comments on zirconium in laterites, and noted that where zircon grains were protected inside pisolites they remained unchanged, but elsewhere zircons "show corroded irregular outlines indicating that they are gradually going into solution." Bramlette (1941) made the same observations with regard to hornblende, pyroxene, and epidote in concretions in sandstone. Frederickson (1948) considers that Zr is more easily removed in solution than Ti, and that because it does not form secondary Zr minerals, the zirconate ions are adsorbed on the aluminum hydroxide minerals and may substitute for Al in the various aluminous minerals; the ionic potentials are very similar, but the ionic radius of Zr^4 (0.87 A) is much greater than that of Al^3 (0.57 A) or Ti^3 (0.64 A). It could, however, substitute for Fe^2 (0.83 A). The movement of silica in bauxite and laterite formation, resulting, if the conditions remain favorable, in its complete removal from a laterite profile, is caused by alkaline solutions, and it would appear that these same conditions are conducive to the corrosion of zircon. It has been noted by the writer that the most strongly corroded zircons are in soil profiles which have been subjected to laterizing conditions, and that the smallest grains have sometimes completely disappeared. It has been suggested elsewhere (Carroll, 1951) that a shorter period of time is required for the complete laterization of a profile developed on granite, than for one on a basic type of rock, as there is actually less silica to be removed (much SiO_2 is in the quartz in granite, and it is considered to be relatively inert). The zircon grains in old lateritic soils are generally strongly corroded (see fig. 2). In a laterite developed on basalt described recently (Carroll and Woof, 1951), only the smallest traces of zircon remained. In a red soil from the Piedmont region of Maryland (Montalto silt loam), the zircons were strongly corroded (fig. 4, a, b, c) in company with corroded zoisite and amphibole.

The finding of zircon with outgrowths by Butterfield (1936) and Smithson (1937), as well as others, indicates that conditions were favorable for the addition of material from solution, probably without much transport. The sandstones in which these zircons occur were formed under deltaic and estuarine conditions, so that salinity was present when the sandstones were laid down. It is probable, however, that the outgrowths on zircon were formed in the sandstones subsequently to deposition, as suggested by Butterfield in his paper (1936). The outgrowths appear to be similar in origin, but not in appearance, to the zircon outgrowths pictured by Tyler *et al* in the Red Clastic series of Minnesota (1940, pl. 4, fig. 2). Arenaceous beds which are leached by later solutions seem a particularly favorable environment for the production of secondary mineral growths; anatase and brookite are common, tourmaline has been recorded by Krynine (1949), and quartz with outgrowths is characteristic of some sandstones. Sand-

TABLE 1.—*Series A zircon leaching experiment*

	Zircon A Australian concentrate[a]	Zircon B Australian concentrate	Zircon C Alluvial, Colorado[b]
Size, mm.	0.25–0.12	0.12–0.06	0.25–0.12
Weight, gm. original	2.0	2.0	2.0
Weight, gm. final	1.11	2.76	1.80
Difference	−0.89	+0.76	−0.20
Test for Zr:			
Alizarin	nil	nil	nil
Cupferron	trace	trace	trace

[a] Commercial concentrate; monazite removed with Franz Isodynamic Separator. Foote Mineral Co. sample donated by Mrs. L. Dryden.

[b] Alluvial concentrate from Lincoln, Placer Co. Colorado; ferromagnesians removed with Franz Isodynamic Separator. Donated by Dr. H. A. Ireland, University of Kansas, Lawrence, Kansas.

stones commonly contain siliceous solutions which cause the hardening of the rock when it is quarried for building material. Solution of a little zirconium and its association with siliceous solutions in the right proportions under stagnant conditions seem to be all that is necessary to produce zircon outgrowths. Alkalinity is required for the solution of zirconium, but acidity for its crystallization as zircon.

EXPERIMENTS ON THE WEATHERABILITY OF ZIRCON

It was thought possible that some results might be obtained by treating zircon grains in various ways to similate simplified natural conditions in sediments and soils, but the results obtained were practically negligible. However, the procedures are recorded here so that the scope and limitations of the experiments can be made known. There is very little in the literature concerning the solubility of zircon, but experiments done by Bouyoucos (1921) indicate that it is slightly soluble, when freshly crushed, in pure water.

Series A.—Three samples of clean zircon grains were packed at three levels in a column of synthetic cation exchange resin, Amberlite I.R. 120, with reactive capacity of 4 m.e. per gm dry weight. The column was 24 mm in diameter and 140 mm long; it contained 22.5 gm Amberlite (dry weight). The column was connected to a reservoir which delivered 1.7 liters of distilled water per day by dripping slowly through. The leaching was continued with the same water for 135 days, so that a total of 229.5 liters was leached through the column. After 80 days the column was washed through with 25° ml 2N HCl, followed by 500 ml water; this was repeated after another 7 days, and the washings were kept. After 135 days the column was washed through with 2 liters of 0.5 per cent oxalic acid which was also kept. The zircon samples were recovered from the Amberlite of the column by panning and separation in bromoform; the Amberlite is inert and the separation was easily made.

The leaching solutions were reduced in volume and tested for the presence of Zr with the alizarin colorimetric test (Sandell, 1950, p. 641); the results were negative. When a few ml of freshly prepared 6 per cent cupferron solution was added to a 50 ml aliquot of the leaching solutions precipitates were formed, but when ignited, the weights were negligible. The details are given in table 1. The brownish precipitate from the leaching with HCl may have been Fe coming from the Amberlite or from the few iron-bearing minerals left in the zircon concentrates. There were no iron coatings on the Australian zircon, but there may have been some on the Colorado zircon concen-

TABLE 2.—*Series B zircon experiment (Australian zircon concentrate)*

No.	Mixture	Weight after 164 days	Zr test with cupferron
(1)	5 gm zircon 0.5 gm fluorite[a] 25 ml distilled water	very slightly reduced	small white ppt.
(2)	5 gm zircon 25 ml distilled water	very slightly reduced	small white ppt.
(3)	5 gm zircon 25 ml 0.01 N HCl	very slightly reduced	pale yellow ppt. wt. 1 mg.
(4)	5 gm zircon 0.5 gm fluorite 25 ml 0.01 N HCl	very slightly reduced	pale yellow ppt.

[a] Finely crushed pure fluorite crystals.

trate. There was a reduction in grain size of Sample A which accounts for the gain in weight in Sample B; also a small quantity of grains passed the 0.06 mm sieve. The total loss in weight of A plus B is 0.13 gm or approximately a 3 per cent loss in weight; this represents 64 mg zirconium (some of this loss in weight may represent removal of odd grains of rutile left in the sample which contains about one grain per hundred of rutile). The loss on C is 10 per cent; representing 100 mg zirconium; this may be partly made up of odd grains of epidote and amphibole which were not removed electromagnetically. The zircon in Sample C is in fresh, euhedral, unabraded grains. Zircon grains were not lost by adhering to the Amberlite or to the glass column.

If the loss in weight in zircon could be related to the leaching solutions by their content of zirconium, which should be present in a weighable quantity, then this experiment would be regarded as highly satisfactory. It is possible that Zr has entered into an exchange relationship with the Amberlite, and that leaching with HCl and oxalic acid failed to remove it. The use of exchange resins sometimes requires a complexing agent for the removal of certain ions, Zr being one of these. There is an extensive literature on the use of synthetic resins for Zr removal (Kunin and Meyers, 1950) but time was not available to test various methods of removal. It is hoped at a later date to repeat this and other experiments with zircon. This experiment can only be regarded as suggestive at the present time.

Microscopic examination showed that the only effect of the treatment which could be recognized was a reduction of the polish on the surface of the grains in A. and B. There was no recognizable effect on the grains in C.

Series B.—Several 5 gm samples of Australian zircon concentrate were placed in 50 ml erlenmeyer flasks with various mixtures, shaken up, and allowed to stand for 164 days. The mixtures and results are as given in table 2.

Microscopically the grains showed no change except for no. (3) in which they appeared to have an increased number of cracks.

Series C.—Several 2 gram samples of Australian zircon concentrate 0.25–0.12 mm diameter were placed in 50 ml erlenmeyer flasks with various mixtures, shaken up, and allowed to stand for 147 days. The mixtures and results are given in table 3.

Microscopically there is no noticeable change in nos. (1) and (3), but in no. (2) additional cracks appear to have developed on the surfaces and ends of grains.

Without repetition and more rigorous control of experimental details these

TABLE 3.—*Series C zircon experiment (Australian zircon concentrate)*

No.	Mixture	Weight after 147 days	Zr loss[a]	Zr test with cupferron
(1)	2 gm zircon 0.5 gm ground lignite 0.5 gm $CaCO_3$ 25 ml distilled water	loss, 0.035 gm	13 mg	small white ppt.
(2)	2 gm zircon 0.5 gm lignite 25 ml distilled water	loss, 0.021 gm	10 mg	very small white ppt.
(3)	2 gm zircon 25 ml 0.3 N $NaHCO_3$	loss, 0.01 gm	5 mg	very small white ppt.

[a] Calculated from weighed loss.

crude experiments cannot be regarded as giving any conclusive results. However minute precipitates were formed with cupferron indicating the presence in solution of Zr, Ti, V, Ta, or Fe. Ti and Fe are the most likely contaminants here. These experiments are merely suggestive of lines which might profitably be followed in examining the processes of corrosion of zircon grains, which undoubtedly does take place particularly in certain soils. Another line of inquiry into the behavior of zirconium under weathering conditions is that of the zirconium in certain amphiboles. On the alteration of these amphiboles the zirconium may go into the circulating solutions; or it may crystallize into some alterations product which has not yet been recognized.

DISCUSSION

Before the zircon or zirconium content of a soil can be safely used as a measure of that soil's development the following observations appear to the pertinent:—

(1) In sedimentary rocks and the soils developed from them, the zircon will be exceedingly stable because all the unstable zircon grains, particularly of type (3) above, will have already been removed, presupposing that the weathering and sedimentation processes have been rigorous, and that zircon was originally present in the distributive province. An example of the probable removal of unstable types of zircon is the Australian commercial zircon produced by concentrating ancient zirconiferous sands. The grains in an average concentrate are all robust, well-rounded, and practically free from inclusions; about one grain in 200 is zoned, and it shows the effect of gradual alteration and removal. Such grains are definitely corroded, and the original sharpness of the crystal faces has been replaced by a more rounded form which is quite different from that produced by abrasive rounding.

(2) It is not yet known with certainty that all igneous and metamorphic rocks have either a constant amount of zircon for any particular rock type, nor constant varietal features. The latter has been proved in several cases (Brammall, 1928; Tyler *et al*, 1940). If there is patchy distribution of zircon either as to quantity or varietal features, then its use as suggested by Haseman and Marshall (1945) is invalidated. The use of varietal features may have greater value than the actual quantity of zircon present.

(3) The weathering and soil-forming processes have an effect on zircon which so far has not been assessed. Frederickson (1948) and Goldich (1938) have noted this, and the writer's collection of heavy residues of soils contains innumerable corroded grains of zircon. The appearance of individual zircon grains under the microscope will give useful information as to the past history of the material in which they occur; for instance, with

loess as a parent material, the zircons will be small, evenly graded, with fresh appearance, and uncorroded, but consideration of the source of the loess is important. If derived from sedimentary rocks the zircon will most likely show some signs of abrasion: a rounded shape, matt or polished surface. If from igneous rocks, the zircon is liable to be in small, euhedral, unabraded grains, and may all be of one or two varieties.

CONCLUSIONS

Observations have shown that zircon grains become corroded, particularly in soils of lateritic type which have been subjected to alkaline leaching for considerable periods of time. All zircon, even that classed as the normal type, is not equally susceptible to corrosion, and reasons are given for this. The literature suggests that zircon may be slightly soluble in Ca-bicarbonate or Na-carbonate. Observations on a number of zircon grains from different localities suggest acid as well as alkaline leaching, as in soil formation, acting through long periods of time, for the production of corrosion and the final removal of zircon. Experiments done with a synthetic cation exchange resin and with various mixtures during a limited time were not conclusive, but should be repeated. The use of zircon as an indicator of changes in a soil profile based solely on the chemical determination of zirconium should be interpreted with caution, but varietal features of the actual zircon grains in a heavy residue from a soil could be used with advantage. Perhaps a combination of chemical and microscopic determinations should be expected to give the most correct and satisfactory results in studies in soil development.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, PP. 117–119
FIG. 1, JUNE, 1953

A NEW ROUNDNESS SCALE FOR SEDIMENTARY PARTICLES[1]

M. C. POWERS
Chesapeake Bay Institute of The Johns Hopkins University, Baltimore, Maryland

ABSTRACT

A new roundness scale is proposed. This scale is defined by six roundness classes. Two sets of clay models were prepared to characterize each roundness class. One set shows a high sphericity, the other a low sphericity. Photographs of these models are used to determine the roundness of a particle by comparison.

The roundness of a particle depends upon the sharpness of the edges and corners. It is independent of the shape. Roundness is a physical property of all sediments, and needs to be described. Where fossils are absent from a sedimentary layer it may prove useful in determining environment of deposition. Roundness gives us some idea of the distance traveled by a particle prior to its deposition, and may also indicate tectonic disturbances. It is probable that roundness may be of some use in the correlation of beds. In order to evaluate the validity of such ideas an accurate method of determining roundness must be adopted using a statistically operable roundness scale. For convenience the method should also be rapid.

Hakon Wadell's method for determining the roundness of a particle (Wadell, 1935) involves dividing the average of the radii of the corners of the grain image by the radius of the maximum inscribed circle. This method is the most accurate yet proposed, but it is far too slow to be used in studying sediments on a large scale.

Russell and Taylor (1937) placed particles in classes based on comparison with photographs of type grains. They presented five grade terms. The class limits for the grade terms are referred to the Wadell method. Sediments are usually assigned by estimation using as criteria the amount of wear of corners and edges. The terms and their corresponding mid-points and class limits are given in Table I. Their class limits were not systematically chosen and the arithmetic means of the intervals were used as mid-points. These values do not provide the smaller subdivisions that are needed in the lower values. Practice has shown that the eye can more readily distinguish differences of roundness when the roundness values are low.

Pettijohn (1949) has modified the ar-

TABLE I.—*Roundness grades*

Grade term	Russell and Taylor		Pettijohn	
	Class Limits	Arithmetic Mid-Point	Class Limits	Geometric Mid-Point
Angular	0 to 0.15	0.075	0 to 0.15	0.125
Subangular	0.15 to 0.30	0.225	0.15 to 0.25	0.200
Subrounded	0.30 to 0.50	0.400	0.25 to 0.40	0.315
Rounded	0.50 to 0.70	0.600	0.40 to 0.60	0.500
Well rounded	0.70 to 1.00	0.850	0.60 to 1.00	0.800

[1] Contribution No. 8 from the Chesapeake Bay Institute. Results of work carried out for the Office of Naval Research of the Navy Department and for the U. S. Navy Hydrographic Office.

TABLE II.—*Roundness grades*

Grade Terms	Class Intervals	Geometric Means
Very angular	0.12–0.17	.14
Angular	0.17–0.25	.21
Subangular	0.25–0.35	.30
Subrounded	0.35–0.49	.41
Rounded	0.49–0.70	.59
Well rounded	0.70–1.00	.84

rangement by using a "geometric" scale, and has rounded off the class limits in order to facilitate memory. (See Table I.) Even this "geometric" scale does not provide small enough divisions in the lower roundness values. Pettijohn differs from Russell and Taylor in that he groups particles into classes by using silhouettes rather than photographs for comparison. The comparison characterizes a three dimensional particle by a two dimensional reference although the three dimensional aspect of small particles is easily observed under the microscope by raising and lowering the focal plane through the full thickness of the particle. A more complete picture of the roundness of a particle is obtained in this manner.

Krumbein (1941) relied entirely on visual comparison of the individual grains with silhouettes to arrive at an average roundness value for the samples. He proposed no class limits, but presented silhouettes for nine different roundness values. This method is rather slow and somewhat tedious. Since his roundness values are very close together, it is often difficult to decide which roundness value to assign to a particle.

If another descriptive term, *Very Angular*, were added to Pettijohn's and Russell and Taylor's set of roundness terms, there would be six roundness classes, and this addition would provide an additional division of the lower classes. As is usual, the addition of another class interval would permit hitherto obscured differences to be revealed by the statistical analysis, and would, perhaps, afford a better interpretation of the history of sediments.

The interval from 1.00 to 0.12 has been divided into six intervals in such a way that the ratio of the upper limit to the lower limit of any interval is 0.7. The characteristic point of each interval is the geometric mean of the class limits. The values appear in Table II. The values given have been rounded off to the nearest hundredth. This scale provides the required increasingly fine divisions at the lower end of the scale. It increases the number of classes which permits a better appraisal of variability and it does away with what is basically an infinite open interval from 0.00 to 0.15 present in the older "geometric" classification. It has been our experience that, with the exception of crystals, particle roundness less

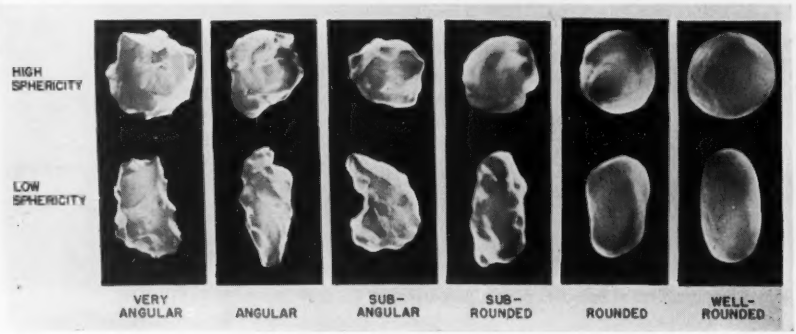

FIG. 1.—Roundness scale.

than 0.12 is not differentiated. The extension of the notion of roundness as defined by Wadell to a crystal is a purely conceptual matter. It is our contention that crystals are not a uniform part of the domain to which the concept of roundness is validly applicable. By deleting crystals and by terminating the division at a point below which observation does not go we have secured a partition into classes that is subject to valid statistical operation.

Figure 1 illustrates particles which fall very near the geometric means of the six class intervals. The particles were modeled from clay to make possible the addition of certain details in the shape, sphericity, and roundness of the particles. Photographs of particles and actual quartz grains whose roundness values were determined using the Wadell method were used as references for the models. Each roundness class is illustrated with two particles, one with a fairly high sphericity, and one with a rather low sphericity. Modeling of the clay and photography of the models was done by Mrs. Josephine B. Thoms.

In determining the roundness of a sample, each particle is assigned to one of the classes depending on the photograph with which it most nearly compares. Fifty or more grains are thus classified by comparison with the photographs. An average roundness for the sample is determined by multiplying the number of particles in each class by the geometric mean of that class and dividing the sum of the products by the total number of particles counted. This method has been used in laboratory work and found to be at least as accurate as other methods proposed.

ACKNOWLEDGMENT

The author is grateful to Dr. Roy L. Ingram of the University of North Carolina for his assistance and criticisms. Blair Kinsman has been instrumental in defining the class limits. His criticisms were most helpful.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, PP. 120–123
FIGS. 1, 2, JUNE, 1953

A NOMOGRAM FOR OBTAINING PER CENT COMPOSITION BY WEIGHT FROM MINERAL-GRAIN COUNTS[1]

ROBERT BERMAN
United States Geological Survey, Washington, D. C.

ABSTRACT

A nomogram for calculating per cent composition from mineral-grain counts reduces several of the more tedious arithmetical operations to two mechanical operations.

The calculation of per cent composition by weight from mineral-grain counts of heavy concentrates is laborious. The basic assumption of grain-counting studies is that the volume of the mineral is proportional to the count; therefore, the samples must be sieved into fractions of uniform size, and the grains of each sieve fraction counted separately. Experience has shown that a single sieve fraction does not necessarily represent the entire sample, for some minerals concentrate in the finer fractions and others in the coarser ones.

Some of the variables in the calculation are:

1. The specific gravity of each of the minerals.
2. The grain count for each mineral in each sieve fraction.
3. The total number of grains counted in each fraction.
4. The weight of each sieve fraction (to determine what portion of the sample it represents).

Accounting for each variable results in a properly weighted factor for each mineral component of each sieve fraction. This factor can be added to the factor for the same component in other fractions, and the total, when prorated to 100, gives the percentage composition by weight of the mineral component in the whole sample.

[1] Publication authorized by the Director, U. S. Geological Survey.

This weighted factor can be found directly, using the nomogram described here. The nomogram has a separate linear scale for each of the mineral components. These scales are calibrated in grain counts, and the size of the divisions is proportional to the specific gravity. The scales are arranged in tandem, with the zero of one scale placed at the count for the previous component on the previous scale.

To illustrate, figure 1 shows the following data: Ilmenite 40 grains, quartz 71 grains, monazite 36 grains. The quartz scale is placed with its zero end at the 40 mark of the ilmenite scale, and the monazite scale is placed with its zero end at the 71 mark of the quartz scale, effecting a simple mechanical addition. On the scales constructed by the author, 1 mm represents one grain of specific gravity 4. These scales are fastened together with paper clips, using a plain card to mark the count of the last component. The total length of the fastened scales now represents the sieved fraction, and the length of each component scale represents the proportion by weight of the indicated mineral component.

The fastened scales are then slid over a set of fan-type variable scales in which one unit varies from 1 to 5 cm. The scales are then placed in such a way that the plain card falls at the per cent of the fraction in the total sample. The per cent of each component can be read off. These are later added to the corresponding

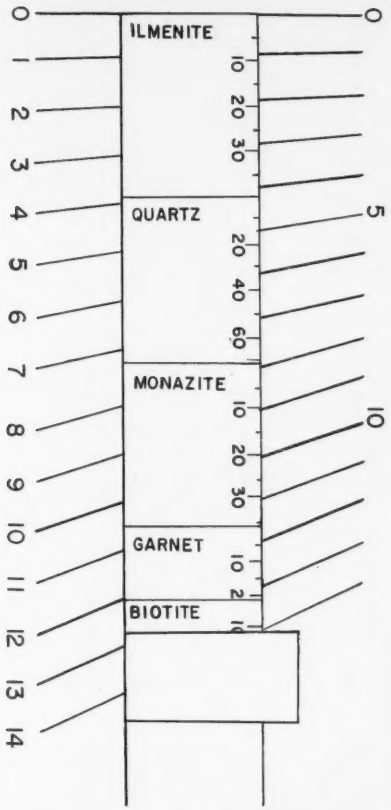

FIG. 1.—Component scales set for grain count of +130 mesh fraction and placed on variable scale so that total equals 14 per cent.

figures in other fractions to obtain the per cent composition by weight of the sample.

For example, let us suppose that the grain count of a panned placer concentrate gives the following figures:

	Component	*Grain counts*
+130 mesh	Ilmenite	40
	Quartz	71
3.7 g	Monazite	36
14 per cent of total sample	Garnet	21
	Biotite	12
130–200 mesh	Ilmenite	35
	Quartz	29
22.6 g	Monazite	47
86 per cent of total sample	Garnet	14
	Zircon	22

The individual scales for each of the five minerals in the +130 mesh fraction are overlapped so that the grain count of each mineral shows on the scale for that mineral. The total length of the scales in this example is about 17 cm. The +130 mesh fraction is 14 per cent of

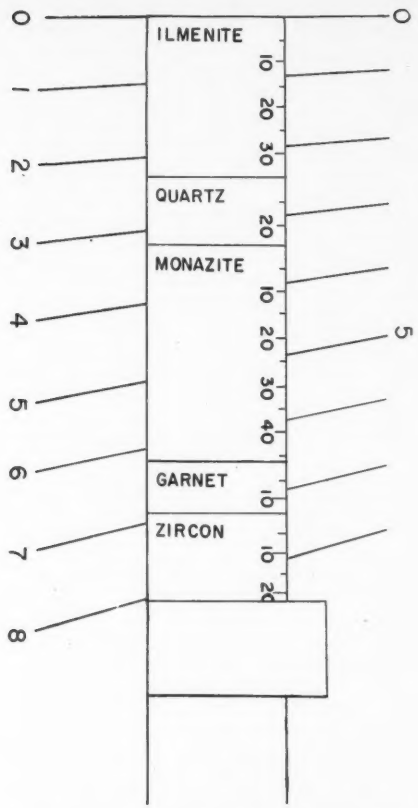

FIG. 2.—Component scales set for grain count of 130–200 mesh fraction and placed on variable scale where total equals 86 per cent.

the sample, and the joined component scales are slid over the variable scale to find a place where they equal 14 units. Figure 1 shows them in the proper position. It can be seen that the ilmenite scale covers 4 units out of the 14 (meaning that 4 per cent of the entire sample consists of ilmenite in the +130 mesh sieve fraction), the quartz scale 4 more, the monazite scale about $3\frac{1}{2}$, the garnet 2, and the biotite about half a unit. A ruler may be used as a guide so that each component scale can be read from the nearest line, or from the zero line of the variable scale.

In arranging the nomogram for the 130–200 mesh fraction (fig. 2) multiply the readings of the variable scale by 10. It is seldom necessary to use any other factor, unless too few grains have been counted to obtain a proper determination. Using the same methods as before, we get the following percentage results:

Ilmenite	23
Quartz	10
Monazite	32
Garnet	7
Zircon	14
	86

Adding the two sets of percentage results:

	+130 mesh	*130–200 mesh*	*Total*
Ilmenite	4	23	27
Quartz	4	10	14
Monazite	4	32	36
Garnet	2	7	9
Zircon		14	14
			100

This nomogram was used in calculating the composition of several hundred samples collected in North Carolina during the summer of 1951 by W. C. Overstreet of the U. S. Geological Survey. We have found that calculation time for each sample is reduced from nearly 20 minutes to approximately 7 minutes. In calculating compositions of a large number of samples the nomogram method reduces mental fatigue and thereby minimizes the number of errors.

This report concerns work done on behalf of the Division of Raw Materials of the U. S. Atomic Energy Commission.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, PP. 124–137
JUNE, 1953

ABSTRACTS OF PAPERS TWENTY-SEVENTH ANNUAL MEETING SOCIETY OF ECONOMIC PALEONTOLOGISTS AND MINERALOGISTS

The Twenty-seventh Annual Meeting of the Society of Economic Paleontologists and Mineralogists was held in Houston, Texas, on March 23–26, 1953. Some of the sessions were held jointly with the American Association of Petroleum Geologists and with the Society of Exploration Geophysicists.

Society officers for the year March, 1953–April, 1954 are as follows:

President: HAROLD N. FISK, Houston, Texas
First Past-President: CLIFFORD C. CHURCH, San Francisco, California
Second Past-President: THOMAS H. PHILPOTT, Shreveport, Louisiana
Vice-President: W. THOMAS ROTHWELL, JR., Long Beach, California
Editor, Journal of Sedimentary Petrology: J. L. HOUGH, University of Illinois, Urbana, Illinois
Editor, Journal of Paleontology: GROVER MURRAY, Louisiana State University, Baton Rouge, Louisiana

Pacific Section officers are as follows:

President: ORVILLE L. BANDY, University of Southern California, Los Angeles, California
Secretary-Treasurer: DANA B. BRAISLIN, JR., Union Oil Company, Compton, California.

JOINT SESSION WITH A.A.P.G. AND S.E.G.

March 24, 9:00 to 12:00 A.M.

PRESIDENTIAL ADDRESS

Presiding for S.E.P.M.: HENRY V. HOWE

FORAMINIFERA, AN EVALUATION

CLIFFORD C. CHURCH
Tidewater Associated Oil Co., San Francisco, California
President, Society of Economic Paleontologists and Mineralogists

The historical background of the use of fossil Foraminifera in the search for oil is briefly reviewed. The city of Houston is pointed out as the center of some of the earliest development. The growth of the science in its relation to oil finding is traced, and some of the early problems are discussed.

In the light of the foregoing history, the present-day position of the science is shown. A consideration of the trend in academic micropaleontology today is indicated to be toward biostratigraphy, oceanography, and ecology, in an effort to better understand the distribution of fossil assemblages.

Some of the more recent mechanical aids in research, such as the Kullenberg bottom coring device, sonic depth recorders, and seismic surveys are mentioned. New methods of sample treatment in applied micropaleontology, new lines of approach involving statistical analysis, and a better appreciation of the value of pelagic Foraminifera in regional correlation, are described as some of the modern advances in the scientific study of fossil and living Foraminifera.

SYMPOSIUM: FINDING ANCIENT SHORELINES[1]

March 24, 2:00 to 5:00 P.M.

Presiding: F. W. ROLSHAUSEN AND E. C. DAPPLES

CONTINENTAL SHELF SEDIMENTS OFF THE WEST COAST OF FLORIDA[2]

HOWARD R. GOULD* AND ROBERT H. STEWART
U. S. Geological Survey, Sarasota, Florida

Studies of the continental shelf sediments along the northern and eastern coasts of the Gulf of Mexico have shown that the sediments off the west coast of central Florida differ from those off the coast of northern Florida and Alabama. Off the west coast of central Florida, the sediments occur in six zones roughly paralleling the coast. The innermost zone has an average width of about 15 miles, and its sediments are composed chiefly of detrital quartz, with phosphorite grains abundant locally. Those sediments of the innermost zone are similar to the sediments of the adjacent beaches and to the sediments carried by rivers entering the Gulf, which suggests that the sediments were derived from these sources. The five outer zones, combined, have an average width of 100 miles and their sediments are largely organic.

Off the coast of northern Florida and Alabama, quartz is the dominant constituent of the sediments for a distance of about 50 miles from shore. The greater seaward extent of detrital sediments in this area probably results from the higher relief of the adjacent landmass and from the entry of larger streams into the Gulf in this region.

The faunal changes off central Florida are about the same as those off northern Florida and Alabama, as determined by O. L. Bandy.

The results of these investigations, when applied to the problem of locating ancient shorelines, seem to warrant the following conclusions:

1. Where the adjacent landmass has low relief and little surface drainage, the detrital marine sediments are limited in extent and quantity, and are mineralogically similar to the beaches and less soluble components of adjacent land areas.
2. Conversely, where the adjacent landmass has moderate relief and large through-flowing streams, the detrital marine sediments are great both in quantity and areal extent but may not be mineralogically similar to adjacent beaches and land areas.
3. Gradations in microfauna are similar in both environments mentioned above, and they should be equally helpful in locating ancient shorelines in both of these environments.

DISTRIBUTION AND ORIGIN OF SEDIMENT UNITS IN THE EASTERN MISSISSIPPI DELTA

P. C. SCRUTON
Scripps Institution of Oceanography, La Jolla, California

The basic factor leading to the development of a river delta is a large excess in rate of supply of sediments over rate of removal. Local variations in the relationship between these rates for different kinds of sediments result in the formation of sedimentary units with different characteristics. The local variations tend to persist laterally and in time, so that three dimensional units are formed. Migrations of the local variations (or environments)

[1] This symposium is to be published in full, with discussion, as Society of Economic Paleontologists and Mineralogists Special Publication No. 3.

[2] Publication authorized by the Director, U. S. Geological Survey.

* Asterisk indicates author who presented the paper.

with delta growth lead to the formation of units of large areal extent which appear to bear predictable relationships to each other. The most important lithologic units which have been recognized in the eastern Mississippi Delta entirely associated with delta growth in order of seaward occurrence are: Marsh Deposits, Delta Front Silts, Pro-Delta Muds, and Offshore Clays. Where other influences do not interfere, these units are deposited approximately concentric to the pass mouths, and their characteristics apparently depend to a large extent on the distance from the source. In addition a sand unit is being deposited north of the active delta through the action of southwest-flowing longshore currents. Another unit, Marginal Muds, occurs between Pro-Delta Muds and the sand and exhibits characteristics transitional between those of the sand and the Pro-Delta Mud. Understanding of the processes of formation and the three-dimensional relationships of the units which result is necessary in order to interpret the stratigraphic column being formed in the delta.

SHORELINES OF THE ATCHAFALAYA REGION, LOUISIANA

WARREN C. THOMPSON
Texas A. and M. College, College Station, Texas

The present shorelines of the world present a great diversity of types. Following the doctrine that "the present is the key to the past," the present coasts can be examined in order to determine how ancient shorelines preserved in the geologic column might be identified. On mountainous coasts shorelines today are marked by sea cliffs which may be preserved. On many low coasts shorelines are marked by extensive coastwise sand deposits in the form of beaches, barrier islands or coastal dunes. On other low coasts the location of the shoreline is indicated only by a thin veneer of marine sediments offshore. In extreme cases, such as on the northeast coast of Florida in the Gulf of Mexico, the shoreline is only faintly marked and identification of a similar shoreline in the geologic column would be extremely difficult.

The present day coast line of the Atchafalaya Bay region is of particular interest because of its anomalous nature. Along the Gulf Coast to the east of Atchafalaya Bay and also westward from Sabine Pass there exist well-developed barrier beaches and islands which could form shoestring reservoirs if preserved in the geologic column. In the Atchafalaya Bay area the trend of the shoreline is marked instead by an extensive oyster reef (*Ostrea virginica*), also a potential reservoir, formed in the brackish water resulting from the Atchafalaya River discharge. Immediately to the westward of the area the coast is being prograded by sediment carried from the river mouth by the predominant westward longshore current. If the present day coast were incorporated in the geologic column the gap in the barrier beach trend that occurs in the Atchafalaya region could reasonably be explained by the proximity of a river carrying predominantly fine sediment. This interpretation would be strengthened by the presence of the oyster reef and also by the occurrence of a broad graduation in the faunal types from sea, across Atchafalaya Bay to land representing respectively marine, brackish and fresh water environments.

SEDIMENT ZONES BORDERING THE BARRIER ISLANDS OF THE CENTRAL TEXAS COAST

FRANCIS P. SHEPARD* AND DAVID G. MOORE
Scripps Institution of Oceanography, La Jolla, California

One phase of American Petroleum Institute Project 51 has been the study of sedimentation along the central Texas coast. The barrier islands of Texas stand as ridges of sand lying between masses of fine sediment on either side. The barrier sands are best sorted on the seaward side and commonly have small but varying amounts of mud in the low, marshy

tracts which border the bays. Numerous characteristics serve to differentiate between the sediments of the bays inside the barriers and the sediments of the open gulf on the outside. In general the bay sediments have more clay than those of the open gulf. This clay content increases away from the barrier islands and is highest in the deeper portions and in the upper bays near the river mouths. Here also lamination of the clays is found but it is absent elsewhere. The bay sediments are also higher in shell content, particularly near oyster reefs. The bay sediments have an abundance of faecal pellets and other indications of the active role of mud-burrowing organisms. Seaward of the barrier islands the sediments change quite abruptly from clean sands to finer sediments which contain alternations between silts, muds, mud-sands, and even zones of sand. There does not appear to be any progressive decrease of grain size on the inner shelf once the line between the sand and the muddy sediments has been passed. These open gulf sediments can be distinguished by their significant content of echinoid spines and glauconite, whereas both of these are rare in the bays. Bay and open gulf sediments can also be distinguished by their Foraminifera, their ostracodes, their macro-organisms, and probably their clay minerals. The characteristics of each of these groups has been studied by a group of specialists who have been cooperating with Project 51.

DYNAMIC GEOLOGY OF THE MODERN COASTAL REGION, NORTHWEST GULF OF MEXICO

ALAN LOHSE

Shell Oil Company, Corpus Christi, Texas

This study is concerned with the relationship of the physiography to the dynamic geology and modern sediments. The area of investigation is a belt along the Gulf coast whose breadth is from the 10-fathom contour in the Gulf of Mexico to the line joining the landward edges of the principal lakes and bays, which extends from Calcasieu Lake, Louisiana, to the south flank of the Rio Grande delta, Mexico. Within this belt of the coastal plain are examples of lagoonal, delatic, tidal flat, littoral and shallow-neritic environments exhibiting freshwater, marine, and euxinic conditions.

Tectonic forces produce the primary features: alternating deltaic and interdeltaic areas. The latter consist of the coastal lake-and-bay regions such as the area which extends from Matagorda Bay to the upper Laguna Madre between the Brazos and Rio Grande deltas. Tertiary counterparts are shale embayments.

The bays, lagoons, barriers, passes, and the sandsheet in southwest Texas are the secondary features. They reflect a delicate adjustment to the physicochemical environment.

The principal agent of the physical environment is the surface wind. In addition to normal wind-work it controls the waves, nontidal currents, and the coastal-drifting processes. These distribute the sediments and shape the secondary features. When analyzing the importance of the wind a distinction must be made between *prevailing* winds, which blow most of the time, and *predominant* winds, which expend the greater amount of energy and usually do the greater amount of work.

PARTICLE-SIZE DISTRIBUTION IN NEARSHORE ENVIRONMENTS

D. L. INMAN* AND T. K. CHAMBERLAIN

Scripps Instutition of Oceanography, La Jolla, California

The patterns of areal variation in the distribution of sediments from several beach and nearshore environments along the California and Gulf of Mexico coasts have been studied. A detailed investigation was conducted on a closely-spaced grid in one of the areas. Sufficient areal and seasonal samples were obtained so that the characteristics of the distribution of particle size with time and distance could be determined. In general it was

found that seasonal changes in the sediments were not of sufficient magnitude to overshadow properties that can be attributed to the location or environment of deposition of the sediment. The investigation also indicated that the most satisfactory technique for studying the relation of particle size to environment utilizes a three dimensional plot of the basic statistical parameters, median diameter, standard deviation measure, and skewness measure.

The findings and techniques from the detailed studied area have been used with success in more complex localities such as the embayments along the Texas Coast and in the Mississippi Delta region. Several types of sediment have been established on a basis of particle-size distribution, and these types have been correlated with the various nearshore environments. For example, the beach sands were found to be among the coarsest and best sorted material in each area, while samples from localities where two distinct sediment types mix, show a pronounced positive skewness and are poorly sorted.

JOINT SESSION WITH A.A.P.G. AND S.E.G.

March 25, 9:00 to 12:00 A.M.

Presiding for S.E.P.M.: CLIFFORD C. CHURCH

APPLICATION OF NEW TECHNIQUES TO BASIC OBJECTIVES IN PALEONTOLOGY

HEINZ A. LOWENSTAM
California Institute of Technology, Pasadena, California

The composition of fossil skeletal remains is becoming the subject of increasing geochemical scrutiny by means of refined tools and newly developed techniques. An example of their application in the paleobiologic inquiry may be seen in Urey's isotopic paleo-temperature method, which yields quantitative determinations of biologically recorded temperatures from oxygen isotopic composition of carbonate skeletons. This focuses attention on the potentials of skeletal building materials as added sources of information regarding the physiologic and ecologic aspects.

Recent studies of mineralogy, trace element concentrations and isotope abundances give promise of new information for analysis of life's past history. The raw materials considered are carbonate skeletons synthesized by marine organisms. The paper examines the methods of approach and their limiting factors in the analysis of the fossil record.

SOME RECENT DEVELOPMENTS IN WELL LOGGING

LOY M. CHARTER
Shell Oil Company, Los Angeles, California

ENVIRONMENTAL INTERPRETATION OF FACIES MAPS

W. C. KRUMBEIN AND L. L. SLOSS
Northwestern University, Evanston, Illinois

GEOPHYSICAL PROGRESS

HENRY CORTES
Magnolia Petroleum Company, Dallas, Texas

AEROMAGNETIC ANOMALIES OVER KNOWN STRUCTURES AND OIL FIELDS

N. C. STEENLAND
Gravity Meter Exploration Company, Houston, Texas

SEISMOGRAMS AS AN AID TO GEOLOGICAL INTERPRETATION IN THE POZA RICA AREA

D. W. ROCKWELL
Geophysical Service, Inc.
AND
ANTONIO GARCIA ROJAS
Petroleos Mexicanos, Mexico, D. F.

GENERAL PALEONTOLOGY
March 26, 2:00 to 5:00 P.M.

Presiding: CHARLES W. STUCKEY, JR. AND ALVA C. ELLISOR

AGE OR BIOFACIES CORRELATION IN PETROLEUM GEOLOGY

W. T. ROTHWELL
Richfield Oil Corp., Long Beach, California

Tertiary basins along the Pacific Coast of North America show lateral faunal changes in the same formation, reflecting differences in such ecologic factors as depth of water, distance from shore, composition of sediment, and abundance of food. Distribution studies of living ostracodes and published ecologic data on foraminifers and fish provide a sound basis for paleoecologic interpretation of fossil faunas encountered in petroleum geology. Fossil fish scale genera serve as useful checks on ecologic environments of the early Tertiary Foraminifera, the species of which are extinct or distantly related to living species. Ecologic "biofacies" zones have been used widely in local California stratigraphic correlations as "time zones," but, viewed regionally, these often prove to be of little or no significance with change in geographic position in a basin. Surface sampling, coring, and close cooperation between paleontologist and field geologist promote detailed data with which well cuttings can be correlated in working out facies problems.

DEVELOPMENT OF UPPER CRETACEOUS FORAMINIFERAL FAUNAS IN COLUMBIA, S. A.

V. PETTERS
International Petroleum (Columbia) Ltd., Bogotá, Columbia

A sequence of seven different foraminiferal faunas is recognized in the seaway which crossed through central and northern Columbia from Ecuador into Venezuela. Their ages range from Lower Turonian (Eagle Ford) to Maestrichtian (Navarro) according to the accompanying ammonite faunas. A uniform, moderately deep sea with badly aerated bottom water is indicated by planktonic assemblages for Turonian (Eagle Ford) and Lower Coniacian (Austin) time. Conditions were favorable for the generation of oil. During late Coniacian (Austin) time, environmental conditions became more diversified due to a progressive shallowing of the sea from south to north. Benthonic faunas appeared at first in the south and spread gradually northward. An inverse faunal movement is noticed for the very end of Coniacian (Low Austin) time. During Campanian (Taylor) and Maestrichtian (Navarro) time, four distinct faunal provinces are recognized from north to south: the Ranchería (outer neritic, towards open sea), the Barco (more endemic fauna), the Upper Magdalena and the Bogotá provinces (restricted shallow sea). Transition zones connect them, with exception of the two latter ones.

A southwest-northeast running threshold area (not necessarily a solid land barrier) is assumed to be responsible for the division between the Upper Magdalena and Bogotá provinces. The trend of the boundary between the Ranchería and Barco facies suggests that the Perija mountains were not emerged from the sea in Upper Cretaceous time, and that the present Ranchería and Cesar valleys originally formed part of the Maracaibo basin.

The Ranchería fauna is considered a southern outpost of Gulf Coastal, and the Upper Magdalena fauna as a northern outpost of Peruvian faunas. The Upper Cretaceous seaway which crossed through Columbia may have connected the otherwise separated "Atlantic" and "Pacific" oceans, without however, permitting more than a local intermingling of the two faunal elements.

The "BARLOW FAUNA," a BIOFACIES OF THE EARLY UPPER CRETACEOUS LOWER ATKINSON FORMATION IN THE SOUTHEAST GULF COAST

ESTHER R. APPLIN
U. S. Geological Survey, Jackson, Mississippi

ZONATION OF THE OCALA GROUP IN PENINSULAR FLORIDA

HARBANS S. PURI
Florida Geological Survey, Tallahassee, Florida

Regional studies of the Ocala limestone justify the use of the term Ocala group, which is redefined to include all calcareous sediments of Jackson age that occur east of Tombigbee River. So defined the group includes the following formations: Inglis, Williston and Crystal River. The Inglis and the Williston were both originally described as members of the Moodys Branch formation, but are here raised to formational rank because of the distinctiveness of their lithology and fauna, and because of their lithologic differences from the type Moodys Branch. The name Crystal River formation is proposed for the 108 feet of limestone exposed in the Crystal River Rock Company quarry, Section 6, Township 19 South, Range 18 East, Citrus County, Florida. It includes all calcareous sediments of upper Eocene age lying stratigraphically between the Williston formation and the overlying Oligocene limestones.

The following faunizones are recognized within the three formations of the Ocala group:

Crystal River formation
- *Planorbulinella* faunizone
- *Nummulites vanderstoki-Hemicythere* faunizone
- *Lepidocyclina-Pseudophragmina* faunizone
- *Spiroloculina* faunizone

Williston formation
- *Nummulites moodybranchensis* faunizone
- *Operculina mariannensis-Nummulites jacksonensis* faunizone

Ingl's formation
- *Periarchus lyelli floridanus* faunizone

RECLASSIFICATION OF THE MIOCENE OF THE FLORIDA PANHANDLE

HARBANS S. PURI
Florida Geological Survey, Tallahassee, Florida

There has been considerable doubt as to the sequence of the various formations in the standard Miocene section of western Florida. This may be attributed to the scattered outcrops, the homogeneity of the sediments, and lack of data on the strike, dip, thickness and structure of these beds. Stratigraphy was based entirely on supposed faunal evolutions, disregarding lithofacies, biofacies, lithotopes and biotopes.

The present study embraces Okaloosa, Walton, Holmes, Washington, Bay , Leon, Jackson, Calhoun, Gulf, Liberty, Franklin, Gadsden, and Wakulla counties. Samples from 58 outcrops and 20 auger holes and water wells were studied Structure sections and faunas of the Miocene indicate the presence of a number of lithofacies and biofacies, which are a measure of recurrence of similar conditions, and are reflected in both the lithology and fauna. Similar depositional types are considered as magnafacies, while dissimilar components within them are designated as parvafacies These patterns, resulting from shifts in time and space distribution of environments, coincide with formational boundaries. Three magnafacies are recognized (Tampa, Chipola and Choctowhatchee), which include the following parvafacies:

Tampa magnafacies
- "Limy" parvafacies
- "Silty" parvafacies

Chipola magnafacies
Chipola parvafacies
Oak Grove parvafacies
Shoal River parvafacies

Choctawhatchee magnafacies
Yoldia parvafacies
Arca parvafacies
Ecphora parvafacies
Cancellaria parvafacies

THE OSTRACODE GENUS PARACTHEROIS AND ITS ECOLOGICAL IMPLICATIONS IN THE STUDY OF MUDLUMP SEDIMENTS

HAROLD V. ANDERSEN
School of Geology, Louisiana State University, Baton Rouge, Louisiana

The Ostracode genus *Paracytherois* was established by Müller in 1894 while working with the fauna and flora of the Gulf of Naples. The five species described by Müller were associated with calcareous algae. Subsequent studies have extended the reported geographical range of *Paracytherois* from the Coast of Norway to Antarctica, and its purported habitat from 10 fathoms near the British Isles to 385 meters in Antarctica, with the greatest number of occurrences reported from sediments at 45–60 fathoms. It appears, therefore, that the mudlump sediment off the mouth of South Pass of the Mississippi River, which contains the Ostracode genus *Paracytherois*, was originally deposited in an environment comparable to that near the outer edge of the present Continental Shelf.

HURRICANE LENTIL EOCENE FORAMINIFERA

LOUIS DE A. GIMBREDE
Texas A. and M. College, College Station, Texas

Foraminiferal data are presented as evidence that the Hurricane lentil, in the Landrum member of the Cook Mountain formation of the Claiborne Eocene, was deposited during an advance and retreat of the East Texas Eocene sea. Faunas of the lentil indicate marine conditions, while brackish water Foraminifera occur in the beds above and below. The presence of pelagic forms throughout the lentil suggests open sea conditions. Benthonic forms give evidence that water depth never exceeded 100 fathoms.

A range chart showing the abundance of 127 species of Foraminifera indicates that three faunal zones exist, coinciding with the lithologic facies of the lentil. Approximately 30 species have been selected and plotted on an ecologic chart as indicators of depth changes.

ILLUSTRATED REPORT ON THE ALGIERS INTERNATIONAL GEOLOGICAL CONGRESS

M. K. ELIAS
Nebraska Geological Survey, Lincoln, Nebraska

SEDIMENTOLOGY

March 26, 9:00 to 12:00 A.M.

Presiding: SHERMAN A. WENGERD AND W. T. ROTHWELL

FAUNAS, FACIES, AND TECTONICS

L. L. SLOSS* AND W. C. KRUMBEIN
Northwestern University, Evanston, Illinois

The principle of lithologic associations related to environment and tectonics has been investigated and illustrated by a number of workers. Further examination of these associations and their fossil components suggests a parallel grouping of biologic associations. The manner in which these may be analyzed includes the following:

1. Environmental indicators in the biologic complex.

2. Lithofacies, distribution, and isopachs of the associated strata.
3. Interrelationships between lithic and faunal patterns.
4. Evaluation of the relative influence of depositional environment and tectonic state on the observed associations.

Inherent in this approach is the necessity for identification, differentiation, and mapping of units of strata defined by environmental factors. These may coincide with or transect rock and biostratigraphic units.

The problems of biologic associations are illustrated by examples from the Paleozoic and Mesozoic.

SEDIMENTARY FRAMEWORK OF THE MODERN MISSISSIPPI DELTA

H. N. FISK, E. McFARLAN, JR.*
Humble Oil & Refining Co., Houston, Texas
C. R. KOLB
U. S. Waterways Experiment Station, Vicksburg, Mississippi
AND
L. J. WILBERT, JR.
Consulting Geologist, Monroe, Louisiana

The main sedimentary units in the bird-foot delta of the Mississippi, as disclosed by borings, closely correspond to the principal depositional environments. The distributaries (passes) are underlain by "fingers" of fine sand and silt approximately 200 feet thick and 5 miles wide. These "fingers" pinch out laterally into relatively sand-free, brackish water, clay "wedges" which underlie interdistributary troughs. Both sand "fingers" and clay "wedges" grade seaward into pro-delta marine clays and are underlain by a 300-foot thick stratum of these clays.

The growth of the delta is controlled by the progradation of the distributary mouth bars. A cone-like bulge of the Gulf bottom, the submarine extension of the bars, slopes away from the mouth of each distributary. Each cone has approximately the same width as a sand "finger," and the depth of the break in slope at its toe at 120 to 200 feet, approximates the thickness of a "finger." As the bars build seaward, the passes lengthen and create the interdistributary troughs. These fill with clay and fine silt, the topmost layers of which are natural levee and marsh deposits, 5 to 20 feet thick.

Using information from borings, maps dating back to 1839, and hydrographic surveys, it is possible to establish the rate of seaward extension of the Southwest Pass bar. By projecting this rate into the past, it was determined that the present bird-foot delta started to form approximately 400 years ago.

A RATIONAL THEORY OF DELTA FORMATION AS EXEMPLIFIED BY THE PRESENT-DAY MISSISSIPPI DELTA

CHARLES C. BATES
Division of Oceanography, U. S. Navy Hydrographic Office, Washington, D. C.

A study of the oceanographic aspects of delta formation has resulted in the development of a comprehensive theory concerning the inflow of sediment-laden water into or through a currentless, tideless, wave-free basin. Three distinct types of flow are possible, Hyperpycnal inflow (inflow more dense) is represented by the turbidity current and may have been an important factor in eroding the Mississippi Submarine Canyon. Homopycnal inflow (inflow equally dense) under an appreciable hydraulic head explains the flow of Mississippi River water through Lake Ponchartrain, Louisiana when the Bonnet Carre Floodway is operating. This type of flow pattern develops the classical Gilbert-type delta with top-, fore-, and bottom-set beds. Hypopycnal inflow (inflow less dense) follows the flow pattern of a two-dimensional jet issuing from a slot. Formation of a parabolic bar outside such an orifice favors bifurcation or trifurcation of the channel at that point and development of the bird-foot delta. This flow type is shown to exist at major mouths of the Mississippi River. Rota-

tion of the earth deflects the outflow toward the right, creating a strong littoral current across the tip of southward-facing passes. Also, a tendency develops for right-hand banks to extend farther seaward than left-hand banks.

ORIGIN OF THE DEEPKILL SHALE

S. W. LOWMAN AND R. F. BEERS
Rensselaer Polytechnic Institute, Troy, New York

Many papers have been written about the origin of graptolitic shales in general, and the Deepkill shale in particular. Recent work with Deepkill sediments at the type locality near Troy, New York demonstrates the flow character of a large part of the type section. Early work is reviewed, and current investigation is summarized.

NATURE AND ORIGIN OF AN OLIGOCENE SANDSTONE RESERVOIR

R. H. NANZ
Shell Oil Co., Houston, Texas

The character, distribution, and probable genesis of a typical lenticular oil-bearing sand body of the Gulf Coast Tertiary, namely Zone 19B, an Oligocene sand of the Seeligson Field, Texas, was determined for the purpose of developing and defining criteria by which such sand bodies can be readily identified and delimited in the subsurface.

The stratigraphic framework was specified by cross-sections, fence diagrams, isopach maps, and facies maps, prepared from electric logs and core descriptions, and the sediments were characterized by size, shape, thin-section, X-ray, differential thermal, and chemical analyses.

Principal results are: (1) the isopachal trend of the sand body is nearly normal to the regional depositional strike; (2) the sand grades from medium grain at the base to fine at the top and can be divided into three zones on maximum grain size; (3) a typical specimen is a mixture of (a) well-sorted sand, the framework size fraction, (b) interstitial clay, the tail of the size distribution curve, and (c) interstitial calcite cement; (4) the mineral composition is similar to that of a modern river sand, but dissimilar to that of beaches, of the same geologic province.

The sand body is probably an Oligocene alluvial-plain deposit as indicated by its similarity to such Recent deposits in external form, vertical textural zonation, colors of associated shales, mineralogy and presence of mud fragments.

DOWN-WIND CLASTIC LIMESTONES AS A KEY TO THE LOCATION OF BURIED REEF KNOLLS

JOHN W. HARRINGTON, JAMES K. LILES, EDWARD L. HAZELWOOD, LANE E. HORSTMAN, HOWARD E. JACKS AND NEAL NEECE, JR.
Lone Star Producing Company, Dallas, Texas

The Hulldale Field in Schleicher County, Texas, occupies a typical Strawn reef knoll on the East Side Shelf of the Midland Basin. A limestone bed tangent to the knoll extends with an obtuse delta shape from this point many miles to the north, east and south. This bed is believed to be a clastic limestone formed after reef growth time and laid across the then filled back reef lagoon. The hypothesized origin is that wave and current action spread a down-wind detritus shoreward towards the Concho Arch. Then with this as a growth platform biogenetic development added to its thickness. This is requisite because the total volume is many times that of the knoll to which it is attached. Similar structures are illustrated from the Recent reefs of Australia and a Mississippian reef knoll near Alamagordo, New Mexico.

The importance of the concept is that by recognizing this structure in the subsurface, the existence and up-wind position of an attached knoll are implied. Paleo-wind vectors may perhaps be obtained by analysis of the lateral distribution and thickness of the layer. Caution is urged in employing these criteria because the key beds are known to extend for as much as twenty miles. Theoret-

ically there might be only one attached knoll and this should be at the most windward point.

The paper is an illustration that paleoenvironment studies will play increasingly important roles in future petroleum exploration.

SUBAQUEOUS LANDSLIDES AND TURBIDITY CURRENTS IN THE PERMIAN OF WEST TEXAS

J. K. RIGBY
Humble Oil and Refining Co., Houston, Texas

Very coarse conglomerates in the lower Leonard to Middle Capitan section in the Guadalupe, Apache and Chinati Mountains appear to have been formed by subaquous landslides generated on the steep talus slopes of reefs. At least two of these landslides can be traced laterally into rocks showing evidence of deposition by turbidity currents, i.e., progressive sorting away from the source, channeling of underlying beds, and remarkable graded bedding. In one case graded bedding has been observed to extend throughout a conglomerate bed 14 feet thick. The conglomerates are very coarse with numerous boulders up to 15 feet in diameter. One transported block over 900 feet long and 40 feet thick was observed.

MINERALOGY AND SEDIMENTARY PETROLOGY

March 26, 2:00 to 5:00 P.M.

Presiding: W. C. KRUMBEIN AND GORDON RITTENHOUSE

PETROPHYSICS OF THE ST. PETER-SIMPSON FORMATIONS

E. C. DAPPLES* AND H. A. SLACK
Northwestern University, Evanston, Illinois

Sedimentary accumulation within stable cratonic areas commonly includes widespread or "blanket" sheets of sandstone typical of the "pure quartz" type. Such thin sheets of sand are important sources of information bearing on the depositional hydrodynamics, the tectonic condition and nature of the source. In the United States the best known example is the St. Peter sandstone (Chazy and Blackriveran, Ordovician). A knowledge of the characteristics of such widespread sand sheets is important in isolating the factors considered responsible for their occurrence.

The writers have developed the physical characteristics of a typical "blanket" sand from a series of maps including isopachs, lithofacies, packing and cementation (formation factor) and particle size ranges. The body of sandstone studied is restricted to the St. Peter and Simpson formations ranging from localities where sandstone dominates the lithology to the marginal areas to the west, south, and east where the limits of sandstone deposition merge into shales and carbonates. Features discussed are:

a. Areal dimensions of the sheet and consideration of the probable limiting size of such blankets.
b. Volumetric assay of the sand sheet.
c. Variation of formation factor as determined from resistivities given by electric logs.
d. Postulated physical conditions of deposition with the progress of time.

DIAGENESIS OF SOME PENNSYLVANIAN LIMESTONES OF NEW MEXICO

RAYMOND SIDWELL AND G. F. WARN
Texas Technological College, Lubbock, Texas

The Pennsylvanian sediments in Upper Pecos Valley, New Mexico, represent two formations: Sandia (lower) and Madera (upper). They constitute the Magdalena

group. The Sandia has been divided into a Lower Limestone member and an Upper Clastic member. The Madera has been divided into a Lower Gray Limestone member and an Upper Arkosic member. Four tectono-environmental conditions are represented. Each environment corresponds to a distinct member: the Lower Limestone member of Sandia represents a borderline epineritic-infraneritic environment on an unstable shelf; the Upper Clastic member of Sandia represents the fluvial-lacustrine-eolian environment on an unstable shelf; the Gray Limestone member of Madera represents the infraneritic environment on a slightly unstable shelf; the Arkosic Limestone member of the Madera represents a transitional environment on an unstable shelf bordering actively rising highlands.

Texturally, three varieties of limestone are distinguished: 1. microcrystalline or granular, more than 50 per cent of the rock granular; 2. crystalline, composed chiefly of sutured anhedra; 3, fragmental, composed mainly of calcite fragments, quartz and organic remains. Diagenesis in limestone units is highly localized. Some units are practically bypassed, others are congenial hosts. Limestones especially susceptible to metasomatic replacement include the thin- to medium-bedded crystalline and fragmental varieties. Thin-sections and photomicrographs show the diagenesis of various constituents of the limestone follows a definite sequence: 1. for carbonates, aragonite (or calcite) to dolomite; 2. for silica, opal to chalcedony to quartz. Many of the granular limestones result from repeated secondary growths of calcium carbonate on microscopic grains. Granular limestone may become crystalline, and eventually alter to dolomite. Fragmental limestones are abundant in the Upper Madera. They frequently show secondary growths and partial to total replacement; hence they may become partly crystalline in texture. Diagenesis tends to increase porosity in the carbonates and to decrease it in the silicates.

DIAGENESIS OF SOME PENNSYLVANIAN SANDSTONES OF NEW MEXICO

G. F. WARN AND RAYMOND SIDWELL
Texas Technological College,
Lubbock, Texas

Sandstones are represented in each of the four tectono-environments of the Pennsylvanian. The Lower Limestone member of the Sandia formation (borderline epineritic-infraneritic environment) contains two sandstone units, basal and medial. They are gray to greenish-gray, associated with arenaceous limestones and are less than 10 feet thick. The mineral content includes well rounded quartz, chert fragments, partly rounded mica flakes and a few grains of highly altered feldspar. The sandstones of the Clastic member of the Sandia (fluvio-lacustrine-eolian environment) are of two varieties, subgraywacke and quartzose. They are varicolored, medium-bedded to massive, and are less than 20 feet thick. The mineral content includes an abundance of angular quartz, hexagonal mica flakes, a minor amount of partly altered feldspar and a few fragments of phyllite. Locally, some of the sandstone units are associated with coal cyclothems. Sandstones in the Gray Limestone member of Madera formation (infraneritic environment) are chiefly subgraywacke. They are gray to brownish-gray, medium-bedded, and are associated with a sandstone-shale-limestone sequence. The minerals in order of abundance are angular to subrounded quartz, hexagonal flakes of muscovite and biotite, and a few grains of feldspar. The sandstones in the Arkosic member of Madera (transitional environment) are red to green, lenticular poorly sorted, arkosic and conglomeritic. They are associated with red and green shales, micaceous silts and fragmental limestones.

Feldspars, micas, especially biotite, and silica minerals are susceptible to metasomatic alteration. The feldspars, chiefly plagioclase and orthoclase, alter to and are replaced by irregular aggregates of calcite, small transparent mica

flakes (probably sericite), fibrous silica and kaolinite. The iron-magnesium mica is commonly bleached and further disintegrates into kaolinite-like scales. Diagenetic changes in silica minerals include over-growths on quartz grains and alteration from opal to chalcedony to quartz.

THE DIRECTION OF SEDIMENT TRANSPORT OF THE BASAL PENNSYLVANIAN SAND STONES OF INDIANA AND SOUTHERN ILLINOIS[3]

PAUL EDWIN POTTER*
Illinois State Geological Survey,
Urbana, Illinois
AND
JERRY S. OLSON
Connecticut Agricultural Experiment Station,
New Haven, Conn.

The basal Pennsylvanian Mansfield and Caseyville sandstones of Indiana and southern Illinois contain abundant parallel and wedge torrential cross-bedding. Because the total number of possible measurements is so large as to preclude study of the entire population, because the sampling could be reduced to routine, and because the environment was clearly delineated, statistical sampling was appropriate. The sampling scheme assigned variability to five levels:

1. Duplicate planes within the same sedimentation unit.
2. One or more sedimentation units at an outcrop.
3. One to two outcrops in the same or adjacent sections.
4. Several sections in the same township or range.
5. A series of townships or ranges in the outcrop belt.

Since the maximum dip direction of parallel and wedge torrential crossbedding is highly statistically correlated both within and between the sedimentation units of single outcrops, maximum field effort was expended on outcrops, sections, and townships or ranges. Restricting primary attention to the value of cross-bedding as an indicator of the direction of sediment transport, 558 measurements of cross-bedding were obtained from 164 outcrops distributed along 260 miles of outcrop. These data indicate that the Mansfield sandstone and the investigated portions of the Caseyville sandstones had a north-northeasterly source. Given a moderate amount of stratigraphic information, cross-bedding studies of this kind, supplemented by intensive petrologic studies, hold promise of providing inferences concerning regional sediment sources at a fraction of the cost of normal stratigraphic methods.

[3] Published by permission of the Chief, Illinois State Geological Survey.

PETROGENESIS OF LOWER PENNSYLVANIAN SAND STONES OF THE EASTERN INTERIOR BASIN[4]

RAYMOND SIEVER
Illinois State Geological Survey,
Urbana, Illinois

The key to the origin of the sandstones of the Caseyville and Tradewater groups (together formerly the "Pottsville") lies in the detailed quantitative study of three parameters; (1) mineral composition, (2) textures, and (3) stratigraphic variation. Samples have been collected from outcrops and drill holes in Illinois, Missouri, Iowa, Indiana, and eastern Kentucky. Petrographic examination included thin-section study, heavy mineral analysis, and clay mineral analysis. In addition, numerous subsurface stratigraphic sections from all parts of the Basin have been studied.

The spatial distribution of the various facies of these rocks is genetically related to the tectonic sub-provinces of the Basin such as the deeply subsiding central portion in southern Illinois, the relatively stable marginal areas in western and northern Illinois, and smaller areas delimited by major flexures in the Basin, the largest of which is the La Salle anti-

[4] Published by permission of the Chief, Illinois State Geological Survey.

cline. On a smaller scale there is extreme lateral variation in thickness and gross lithology of beds, which is primarily related to the physical characteristics of the environment of deposition and the transporting agent, and secondarily to minor local control of sedimentation by smaller structures in the Basin. Vertical facies changes, as well as the mineral composition of the sandstones, are primarily related to the over-all tectonic control, in particular to the tectonics of the source areas.

A wide variety of sandstone types is present in these two stratigraphic groups. They range from clean orthoquartzites to "washed" subgraywackes and true subgraywackes. They include some conglomerates containing well rounded quartz pebbles and some containing poorly rounded locally derived material. The general mineral composition of the sandstones and, in particular, the varietal types of detrital quartz are related to the composition and degree of tectonism of the source areas.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, P. 138
JUNE, 1953

ANNOUNCEMENTS

GEOLOGICAL ABSTRACTS TO BE PUBLISHED BY AMERICAN GEOLOGICAL INSTITUTE

The American Geological Institute through the good offices of its member societies and the editors of geological journals is beginning a compilation of geological abstracts. The first issue will cover January to June 1953, and will be circulated to subscribers about July 15. Thereafter the issue will be quarterly scheduled as March 30, June 30, September 30, and December 30, of each year.

The subscription price will not exceed $2.00 per year. So that the editors may anticipate printing requirements, will all geologists interested in subscribing to the new service send a postal card at once to: American Geological Institute, 2101 Constitution Avenue, N.W., Washington 25, D. C. *Do not send any subscription money now!*

THE MIDWEST INTER-LIBRARY CENTER CONFERENCE ON GEOLOGIC REFERENCE COLLECTION

An all-day meeting of geologists, representing 13 midwestern university geology departments, was held on February 28, 1953 at the Midwest Inter-Library Center in Chicago. The meeting was called by the Center to explore possibilities for the cooperative acquisition of less-used but important library materials in the field. As an outgrowth of the discussions, the geologists decided that the universities that participate in the Midwest Center might appropriately supplement their own collections with a comprehensive collection of foreign geologic maps and with files of minor and less-used journals, particularly those of foreign academies. Such materials would be acquired by and housed at the Center for the joint use of the several institutions.

The Center purchases less-used research books and periodicals that its individual member libraries cannot afford and extends the use of them freely within the group of participants. It was established in 1949 with a million-dollar foundation grant and is supported and controlled by 16 member universities. Its new library building in Chicago has a three-million-volume capacity bookstacks and is connected with each member library by teletype.

The participants at the February 28 meeting in Chicago were:

Eugene N. Cameron (University of Wisconsin)
Leland Horberg (University of Chicago)
Arthur L. Howland (Northwestern University)
Brian H. Mason (Indiana University)
Archie J. MacAlpin (University of Notre Dame)
R. C. Moore (University of Kansas)
Willard H. Parsons (Wayne University)
John L. Rich (University of Cincinnati)
Charles H. Summerson (Ohio State Univ.)
George A. Thiel (University of Minnesota)
James W. Trow (Michigan State College)
Sherwood D. Tuttle (State Univ. of Iowa)
George W. White (University of Illinois)

Library Consultants were:

Lyle E. Bamber (Natural History Librarian, University of Illinois)
Leon Marshak (Research Staff, John Crerar Library)
George R. Meluch (Purdue University Library)
Mrs. Marie K. Shaw (Science Librarian, Wayne University)

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 23, NO. 2, PP. 139, 140
JUNE, 1953

REVIEWS

Sedimentary Petrography by Henry B. Milner, 3rd ed., reprint, 1952. Pp. xix +666; figs. 100; pls. 52; 5½×8½ ins., cloth. Thomas Murby and Co., 40 Museum St., W.C. 1, London; Interscience Publishers, Inc., 250 Fifth Ave., New York. Price, $9.50.

The third edition of Milner's book is available, as a reprint, after being unavailable for several years. A revision to include the advances made in the field during the past twelve years would have been very desirable, of course, but this worthy text and reference book is still an essential part of the sedimentary petrologist's library. Consisting of fifteen chapters and eight appendices, *Sedimentary Petrography* covers the field from sampling through various laboratory techniques (including mechanical analysis, x-ray, spectrum, fluorescence, microchemical and optical methods) to a consideration of paleogeographic problems. Examples are given of the application of sedimentary petrography to stratigraphic correlation, the study of soils, highway construction, legal and medical problems, and various mineral industries.

A large section giving the diagnostic properties of detrital minerals, which is excellently illustrated with numerous plates, is unsurpassed in any other work, and this section alone makes Milner's book indispensible.

J. L. HOUGH

Einführung in die Kristalloptic by Eberhard Buchwald, 4th ed., 1952. Pp. 138; figs. 121; 4×6 ins., paper. Walter de Gruyter and Co., Berlin. Price, D.M. 2.40.

This little elementary introduction to crystal optics could be likened to a "pocket-book" textbook. It is divided into four parts: I. Double refraction and polarization; II. Interference figures; III. Rotary (polarized) and absorbent crystals; IV. Lattice optics of the visible spectrum. The book is interestingly written in a clear, very readable style on a level of the senior student in geology. The author develops the usual theory given in most English texts on optical crystallography, and then goes farther in considering crystal optics from the standpoint of elementary, non-mathematical, electromagnetic theory. This serves to explain many phenomena that are seen by the student for which he previously had no logical explanation. An English translation would provide an excellent supplementary text for the student in optical mineralogy who wants to go beyond the "how" of things. The last two parts should provide interesting reading for the petrologist with a little German in his grasp.

The paper and printing are on a par with, and perhaps a little more readable than, the "Penguin" series of pocket-books. Numerous line drawings and several photographic reproductions aid in the explanation of the text and are clearly reproduced. The price is very reasonable (about $0.60).

J. P. WEHRENBERG
University of Illinois

Minerals and the Microscope by H. G. Smith, 4th ed., reprinted 1949. Pp. xiv +130; figs. 34; pls. XII; 4¾×7¼ ins., cloth. Thomas Murby and Co., 40 Museum St., London; The Macmillan Co., 60 Fifth Ave., New York. Price, $1.10.

Considering the space available in a pocket-sized book, *Minerals and the Microscope* gives an admirable exposition of petrographic methods and of optical mineralogy. The descriptions of common rock-forming minerals are illustrated with many excellent photomicrographs of minerals in thin-section, showing characteristic habit and features such as zon-

ing. "The study of rocks," which is contained in the last 45 pages, is confined mainly to the igneous rocks. An introduction to fundamental concepts of igneous petrology is presented in a non-dogmatic and easily readable style. This section was reasonably up to date when the fourth edition was first published in 1940. The short section dealing with sedimentary rocks is disappointing, however. It amounts to little more than a superficial statement that the minerals may be identified with the microscope. In a brief discussion of sedimentary petrology the impression is given that the mineral suite of a sediment is determined solely by physical abrasion and by sorting in "water action." The final section on metamorphic rocks, also brief, is another superficial treatment and it accomplishes little more than an introduction to terminology. The modern concept of regional metamorphic zones is not mentioned.

J. L. HOUGH

Practical Microscopy by L. C. Martin and B. K. Johnson, 2nd ed., 1951. Pp. 124; figs. 89; $4\frac{3}{4}\times7\frac{1}{4}$ ins., cloth. Printed in Great Britain. Chemical Publishing Co., Inc., 26 Court St., Brooklyn, N. Y. Price, $2.50.

Authors Martin and Johnson, of the Imperial College of Science and Technology, London, present a small handbook on the principles and methods of microscopy which should be of value to any serious student who employs magnification in his work. The principles of magnification and the functions of the parts of an optical microscope are clearly set forth in the first four chapters. Methods of illumination of the object are treated in detail, including the use of dark-ground and phase-contrast illumination, in chapters five and six. The following eight chapters discuss photomicrography, the metallurgical microscope, preparation of specimens, binocular microscopes, the polarizing microscope, ultraviolet microscopy, interpretation of the image in the microscope, and give a brief exposition of the technique and limitations of electron microscopy. These later chapters are not exhaustive, but they summarize the fundamentals of various specialized fields of microscopy in a clear, easily readable fashion. For example, the chapter on polarized light describes and explains the use of the petrographic microscope, through operations such as determination of optical sign and birefringence to the production of interference figures, but it gives no illustrations of the interference figures. References to more advanced treatments of the various fields are given throughout the text. The numerous line drawings and photographs are excellent.

J. L. HOUGH

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